PB86-122587

Research on Diffusion in Atmospheric Boundary Layers: A Position Paper on Status and Needs

(U.S.) Environmental Protection Agency
Research Triangle Park, NC

Nov 85

U.S. Department of Commerce
National Technical Information Service
NTIS

PB86-122587

EPA/600/3-85/072
November 1985

RESEARCH ON DIFFUSION IN ATMOSPHERIC BOUNDARY LAYERS:

A POSITION PAPER ON STATUS AND NEEDS

by

Gary A. Briggs
and
Francis S. Binkowski
Meteorology and Assessment Division
Atmospheric Sciences Research Laboratory
U.S. Environmental Protection Agency
Research Triangle Park, North Carolina 27711

ATMOSPHERIC SCIENCES RESEARCH LABORATORY
OFFICE OF RESEARCH AND DEVELOPMENT
U.S. ENVIRONMENTAL PROTECTION AGENCY
RESEARCH TRIANGLE PARK, NORTH CAROLINA 27711

REPRODUCED BY
NATIONAL TECHNICAL
INFORMATION SERVICE
U.S. DEPARTMENT OF COMMERCE
SPRINGFIELD, VA. 22161

TECHNICAL REPORT DATA
(Please read Instructions on the reverse before completing)

1. REPORT NO. EPA/600/3-85/072	**2.**	**3. RECIPIENT'S ACCESSION NO.** PB86 122587/AS
4. TITLE AND SUBTITLE RESEARCH ON DIFFUSION IN ATMOSPHERIC BOUNDARY LAYERS: A POSITION PAPER ON STATUS AND NEEDS		**5. REPORT DATE** November 1985 **6. PERFORMING ORGANIZATION CODE**
7. AUTHOR(S) Gary A. Briggs and Francis S. Binkowski		**8. PERFORMING ORGANIZATION REPORT NO.**
9. PERFORMING ORGANIZATION NAME AND ADDRESS (Same as Block 12)		**10. PROGRAM ELEMENT NO.** CDTA1D/03 -1361 (FY-85) **11. CONTRACT/GRANT NO.**
12. SPONSORING AGENCY NAME AND ADDRESS Atmospheric Sciences Research Laboratory - RTP, NC Office of Research and Development U.S. Environmental Protection Agency Research Triangle Park, NC 27711		**13. TYPE OF REPORT AND PERIOD COVERED** **14. SPONSORING AGENCY CODE** EPA/600/09

15. SUPPLEMENTARY NOTES

16. ABSTRACT

The introduction of a new understanding of atmospheric boundary layers (ABLs) has caused a major change in the view of the diffusion of pollutants. The turbulence parameters now standard in ABL work, have provided a method for systematically organizing diffusion parameters. Concurrently with these advances, alternatives to the operational models have emerged, but existing experimental data sets are inadequate for model comparisons and evaluations. The most important knowledge gap is the lack of an adequate specification of the relevant meteorology both at the point of release and downwind. A second major inadequacy is experimental measurements of plume characteristics up to 100 km from the release point. There is also a great need for formulating new operational based models upon this newly acquired experimental data and the new alternative approaches. Finally, it is recognized that a modest but steady effort is necessary.

17. KEY WORDS AND DOCUMENT ANALYSIS

a. DESCRIPTORS	b. IDENTIFIERS/OPEN ENDED TERMS	c. COSATI Field/Group

18. DISTRIBUTION STATEMENT RELEASE TO PUBLIC	**19. SECURITY CLASS** *(This Report)* UNCLASSIFIED	**21. NO. OF PAGES** 250
	20. SECURITY CLASS *(This page)* UNCLASSIFIED	**22. PRICE**

EPA Form 2220-1 (Rev. 4-77) PREVIOUS EDITION IS OBSOLETE

NOTICE

The information in this document has been funded by the United States Environmental Protection Agency. It has been subject to the Agency's peer and administrative review, and it has been approved for publication as an EPA document. Mention of trade names or commercial products does not constitute endorsement or recommendation for use.

The authors, Gary A. Briggs and Francis S. Binkowski, are on assignment to the Atmospheric Sciences Research Laboratory, U.S. Environmental Protection Agency, from the National Oceanic and Atmospheric Administration (NOAA), U.S. Department of Commerce.

ABSTRACT

The introduction of a new understanding of atmospheric boundary layers (ABLs) has caused a major change in the view of the diffusion of pollutants. The turbulence parameters now standard in ABL work, have provided a method for systematically organizing diffusion parameters. Concurrently with these advances, alternatives to the operational models have emerged, but existing experimental data sets are inadequate for model comparisons and evaluations. The most important knowledge gap is the lack of an adequate specification of the relevant meteorology both at the point of release and downwind. A second major inadequacy is experimental measurements of plume characteristics up to 100 km from the release point. There is also a great need for formulating new operational based models upon this newly acquired experimental data and the new alternative approaches. Finally, it is recognized that a modest but steady effort is necessary.

EXECUTIVE SUMMARY

INTRODUCTION

The purpose of this study is to review the current state of science concerning diffusion within the atmospheric boundary layer (ABL) and to identify major issues and research needs. The study is limited to considering only the diffusion of non-buoyant, conservative substances over homogeneous and level or moderately hilly terrain. Thus, chemical transformation, wet and dry deposition, buoyancy effects, and diffusion in complex terrain are all topics which are excluded. Certainly these are very important topics both from the point of view of the research scientist and from that of a regulator; however, each of these topics deserve a review study of the same scope as attempted here. The topics that will be included here are: a review of diffusion experiments and models, and a discussion of research needs. We begin, however, with a highly simplified description of the major meteorological factors affecting diffusion. The main report contains the necessary details and references.

METEOROLOGICAL FACTORS CONTROLLING DIFFUSION

Material released from the ground or from a stack is mixed and transported by air motion. This process dilutes the material and carries it horizontally away from the source. This mixing process is called turbulent diffusion, or simply diffusion. The turbulence responsible for this mixing may result from the wind blowing over the ground, generating turbulent eddies, or it may result from heated parcels of air or thermals that are generated in air heated by contact with ground which has been warmed by the sun. Thus, we distinguish between "mechanical" and "convective" turbulence, because in addition to having different generative mechanisms, they have considerably different diffusive properties.

The current view of the ABL divides it into three conceptual states, the neutral boundary layer (NBL), the convective boundary layer (CBL), and the stable boundary layer (SBL). The NBL occurs when there is a negligible transfer of heat from the ground surface to the air. This occurs with strong winds and usually when most of the sky is covered with clouds. Only mechanical turbulence is generated in the NBL which extends up to or even into the cloud bases. The SBL grows when the ground is substantially cooler than the air above it causing a stable profile to develop. This means that an upward moving parcel of air, for example, experiences a downward restoring force because it is colder, therefore denser, than its surroundings. Mechanical turbulence is generated as the wind blows over the ground, but vertical motions are very much restricted by the stability. Vertical diffusion, consequently, is restricted and plumes of effluent dilute slowly during transport downwind. The depth of turbulent mixing is also very restricted in the SBL, being of the order of 100 m or less. It is under these conditions that some of the most serious air pollution episodes can occur. The CBL occurs in the daytime, with low to moderate wind speeds and clear to partly cloudy conditions. Then the ground is warmed and thermals rise from the surface to generate convective turbulence. The diffusion process is dominated by the vertical transfer of material in the thermals and downdrafts. The height to which these thermals rise is called the mixing depth. As a result of this efficient type of mixing, the CBL often has a "well mixed" condition in which the vertical profile of a pollutant becomes constant from a few meters above the surface to the mixing depth.

On a day without many clouds, and with neither fronts nor shorelines nearby, the diurnal course of the ABL is as follows: starting about an hour before sunrise, the ground has cooled to its minimum temperature. The temperature profile is stable but there is some vertically restricted mechanical turbulence from the surface to the top of the SBL. Any plumes within this SBL are very well defined. Any plumes above the SBL have virtually no vertical dilution but may be spread horizontally by meanders caused by large scale horizontal eddies. As the sun comes up, thermals begin to rise from the ground but must work against the stable temperature profile until all the air in the developing CBL is warmed and mixed. The CBL, thus formed, grows until the diminished surface heating in the late afternoon can no longer overturn stable air above the CBL. Typically, the maximum mixing depth is of the order of a kilometer. Any plumes beneath this height released into the previously stable air are mixed downward to the ground, a process referred to as fumigation. Late in the afternoon, the strength of the thermals diminishes. Soon a new stable layer is established at the surface and a new SBL forms. The transitional times around sunrise and sunset are as yet poorly understood. Little is known about the structure of the ABL and even less about diffusion during these periods.

In a typical diurnal cycle, the large changes in stability correspond to large changes in diffusion rates. Some method of indicating the ambient stability is necessary and there are several indices of stability currently used. The index most indicative of the physical processes at the surface is the Obukhov length, L. By convention L is defined to be negative with an upward flux of heat at the surface (unstable conditions). In principle, this length is defined such that at height $z = L$ above the ground, the buoyant production (or destruction) and the mechanical production of turbulent kinetic energy are comparable. Hence, above $z = L$, the turbulent state of the ABL is dominated by buoyancy effects and below $z = L$, the turbulent state of the ABL is dominated by the mechanical overturning caused by the wind blowing over the surface. Profiles of mean wind speed and other meteorological quantities are functions of z/L near the surface.

Besides the Obukhov length, other quantities of importance in describing the turbulence in the ABL are the surface heat flux, the surface friction velocity scale u^*, and the mixing depth. A velocity scale analogous to u^* has been developed for the CBL. Called the convective velocity scale, w^*, it is essentially defined by the surface heat (or buoyancy) flux the depth of the CBL; w^* has proven to characterize the turbulent velocities within the CBL very well. The depth of the CBL divided by L serves as a useful indicator of the relative importance of the thermals as a major transfer mechanism; the larger this ratio is, the more important thermals are to the description of diffusion. As fundamental as these variables are to our understanding of ABL processes, u^* and the surface heat flux are not easily measured; both are needed to determine L using its definition. An alternative index related to L, called the Richardson number, Ri, uses a temperature difference and velocity differences between two heights: in another form known as the bulk Richardson number, only one wind measurement is used. There are simple methods to convert from Ri to Obukhov length. An estimate of the site roughness length, z_0, is necessary with the bulk Ri approach, while the velocity differences approach strongly demands wind sensors that are properly calibrated. With the temperature differences, and at least one wind velocity, the same methods that yield the Obukhov length can be used to obtain u* and the heat flux. The only other essential measurement needed for characterizing ABL turbulence and diffusion is the mixing depth. This can be obtained from good profiles of temperature and humidity. In the case of the SBL, a profile of wind speed is also required. In recent years it has become possible to estimate this depth by using remote sensors such as lidars and acoustic sounders. Lateral diffusion at large distances is enhanced by wind direction changes with height, which are best determined from wind velocity soundings.

EXPERIMENTAL AND THEORETICAL BASES FOR OUR UNDERSTANDING OF DIFFUSION

The current understanding of diffusion processes is based upon results from field experiments, laboratory experiments and theoretical models. Field experiments may be conveniently grouped into two classes, those in which the tracer material is released from very near the surface and those in which the release is elevated. The data from previous experiments has been very useful in formulating our current understanding of the diffusion process but there are major inadequacies. First, the tracers used in most early field studies were not conservative. This means that as a cloud or plume of tracer was transported downwind, the material stuck to the ground or deposited, thus reducing the concentration further downwind by an uncertain amount. The theoretical view of diffusion prevelant at the time of the earlier experiments was that the vertical profile of concentration was Gaussian. If this were true and the tracer were conservative, the standard deviation of the concentration, Sigma-z, is easily obtained from ground samplers alone. When deposition occurs, however, the Sigma-z estimate is biased by an unknown amount to be an overestimate. A second difficulty with these data is that with one notable exception, the Prairie Grass experiment done in 1956, the experiments did not take sufficient meteorological data for either interpretation using contemporary concept or for validation of current diffusion models. Two recent experiments, one in Copenhagen, Denmark and one under EPA sponsorship in Boulder, CO, had extensive meteorological data. The Danish experiment used a conservative tracer, while the EPA sponsored experiment used remote sensing to get concentration profiles at several locations downwind of the source. This experiment, named Convective Diffusion Observed with Remote Sensors (CONDORS), was designed to verify some new insights into CBL diffusion that were gained from numerical experiments and laboratory studies.

Laboratory studies of ABL diffusion are gaining wider acceptance as the scientific community has learned how to simulate ABL processes more faithfully. The major problem faced in a laboratory simulation is to maintain the proper balance of forces and yet attain the required reduction in scale. In strictly neutral conditions, this is not too difficult. One problem is that if the scale reduction factor is very large, viscous effects smooth out turbulent eddies in the laboratory which would not be smoothed in the atmosphere. This is usually overcome by making the bottom of the wind tunnel have a rougher texture than strict scaling would predict. A major difficulty is properly simulating buoyancy effects. Since gravity cannot be easily scaled down on the earth's surface, density differences must be exaggerated. This distorts other aspects of the simulation so that technical trade-offs are necessary. The most effective use of laboratory work then is to examine a particular, but important aspect of the diffusion problem. One such application which illustrates this point is the work done at the EPA Fluid Modeling Facility that showed that a tall slender building required a shorter stack for good plume dilution than would have been predicted by using building height alone.

Experimental work alone is insufficient for an understanding of diffusion. A proper conceptual framework is also necessary. Most current applications of diffusion models are based upon the Gaussian distribution in both vertical and horizontal directions in order to characterize the diffusion. The Gaussian distributions are based upon empirical data collected primarily during near surface releases over relatively flat terrain. Unfortunately, most of the model applications are for situations that were not considered in the experimental situations which provided the empirical data for the formulation of these models. Finally, for diffusion of non-buoyant releases within a CBL, the Gaussian assumption for the vertical distribution of material was recently shown to be incorrect. Current practice has adjusted the necessary parameters so that surface concentrations are nearly correct for the CBL case even though the vertical profile is not. There are however, several modeling methods that have been demonstrated to be superior for particular applications. In the interest of brevity, only three examples will be mentioned. The first is the statistical method which relates the standard deviations of the concentration distribution directly to the observed turbulent intensities of the wind field. Since the statistics of the wind field tend to be fairly uniform in the horizontal direction, this method has the greatest success in characterizing the horizontal concentration distribution for transport over an area of homogeneous land-use. A second approach is to simulate the diffusion as a random walk or "Monte Carlo" process. Recent work in this area has allowed for vertical and horizontal inhomogeneities in the flow field. A disadvantage to this approach is that it generates individual material trajectories and then constructs the distributions directly. For realistic cases, a very large number of trajectories must be calculated. Finally, a third method is numerical simulation of the diffusion process by solution of the relevant governing partial differential equations in an Eulerian framework. There has been great progress made in formulating these models which require a number of simplifying assumptions. For example the laboratory work on surface releases in the CBL inspired a set of numerical simulations that in turn included a prediction about unexpected behavior of an elevated release. The numerical work predicted that the plume centerline for a non-buoyant release would descend to the surface, then rise again. This prediction inspired further laboratory work that verified the finding. The CONDORS experiment was conducted specifically to confirm the numerical and laboratory findings for diffusion in the CBL. Here is an example of how the various methods for studying diffusion interact in the discovery and confirmation of major new findings in diffusion theory. All of the work mentioned in this example was supported by EPA.

NEEDS FOR FUTURE WORK

Before listing the future needs it is imperative to state at the outset that the human and fiscal resources for these endeavors are not trivial. A modest, steady commitment of personnel and funds for an extended period is necessary, as the work described requires a sustained scientific effort.

The first major need is for field data for selected meteorological scenarios for the development, testing and evaluation of diffusion meteorology models. The characterization of those meteorological variables important to the diffusion process is a major impediment to the implementation of new and emerging approaches to diffusion characterization. To meet these needs requires detailed profiles of the relevant variables within the first kilometer or so of the atmosphere. In addition, sufficient measurements are needed to characterize the surface energy balance, the forcing of the wind fields by the large scale migratory pressure systems, the horizontal temperature field and the cloud conditions. These studies should be conducted over a variety of terrain types using a minimum of preparation and personnel. This would allow the study of the meteorological events of interest at the locations of interest. A portable instrumented tower, a tethersonde (a tethered balloon carrying an instrument package), and an acoustic sounder would be the basic tools for such studies. The development of such meteorological field studies would greatly facilitate the transfer of theoretical results into practical operational models useful for decision making.

A second major need is for the comparison, evaluation and construction of new operational models. An interdisciplinary team with skills in meteorology, chemistry, numerical modeling and statistics is required to meet this need. Such a team would evaluate the models, not only with respect to experimental data, but also with respect to the expected uncertainty in the model inputs. These type of comparisons are necessary to make improvements in the operational models.

A third major need is for diffusion experiments, both in the field and in laboratory settings. The laboratory studies are needed to test theoretical results in specific simplified situations that are free of confounding influences. The field data are needed because there is a wide disparity between the flat homogeneous land where the best of the previous field studies have been conducted and the complex woodlands and urban-suburban developments typical of the American landscape. One example of important information to be gained from such field studies is the characterization of the concentration field at 100 km downwind from the source. Current models extrapolate to this distance based upon measurements to only 3 km. Vertical and lateral profiles of concentration even at widely spaced plume transects would improve the current situation considerably. Other examples include the characterization of releases made during the hours of transition, that is near sunrise and sunset, as the behavior of the atmosphere during these periods is very poorly understood. For any of these examples, a good diffusion experiment is of necessity a good meteorological experiment, since a full complement of meteorological measurements is essential to the success of the experiment and to the continued usefulness of the data set.

CONTENTS

Abstract iii
Executive Summary iv
Table and Figures xvii
Abbreviations and Symbols xviii
Acknowledgements xxi

1. Introduction, Purpose and Scope 1

2. Atmospheric Boundary Layers 3
   Important Meteorological Variables and Scales Affecting Diffusion 5
   The Diurnal Cycle and Transitional Periods 13
   Neutral Boundary Layers 26
   Convective Boundary Layers 29
   Stable Boundary Layers 35

3. Diffusion Experiments 47
   Field Experiments 49
   Laboratory Experiments 74

4. Diffusion Modeling as Practiced 87
   The Gaussian Plume - Its Utility and Limitations 87
   Empirical Diffusion Coefficients and Stability Typing . . . 91
   Evaluations of Sigma Curves and Typing Schemes 103

5. Diffusion Modeling Alternatives 109
   Surface Layer Similarity Models 111
   Gradient Transfer and Higher Order Closure Models 115
   Convective Scaling Models 109
   Large Eddy Simulation Models 130
   Statistical Models 134
   Random Perturbation Models 143

6. What Are the Needs? 149
   Experimental Versus Modeling Needs 149
   What Kind of Experiments are Needed? 152
   Meteorological Measurements 156
   Evaluation and Concentration Fluctuation Issues 160
   Stable Boundary Layer Issues 163
   Convective Boundary Layer Issues 168
   Calms and Other Inconvenient Events 171
   Resource Issues 172

7. Implementation of Improved Modeling Practices 175
   Estimating Needed Parameters from More Obtainable Measurements 176
   Building Better Operational Models 180

8. Assessment 183
  Atmospheric Boundary Layers 184
  Diffusion Experiments 186
  Diffusion Modeling as Practiced 190
  Alternative Modeling Approaches 194
  Needs for Diffusion Experiments 200
  Research Modeling Needs 206
  Technical Transfer Needs 207

References 210

TABLES

Number		Page
1	Summary of Tracer Experiments in Homogeneous Terrains	52-53

FIGURES

1	Temporal behavior of measurements in a CBL at the Boulder Atmospheric Observatory, Colorado, USA 13 September 1983	20
2	Temporal behavior of measurements in a SBL at Cabauw, The Netherlands, 30-31 May 1978	23

LIST OF ABBREVIATIONS AND SYMBOLS

ABL	Atmospheric Boundary Layer
b	Dimensionless constant in Eq. 2.5 for h, SBL
C	Concentration (generalized)
CBL	Convective Boundary Layer
D	Wind direction
f	Coriolis parameter
g	Gravitational acceleration
h	Mixed depth height for NBL or SBL
H*	Turbulent flux of buoyant acceleration at the surface: $H^* = (g/T_v')\,\overline{w'T_v'}$
i	Subscript, "inversion"
k	von Karman's constant, ≃ 0.4
K	Eddy diffusivity (generalized) (K_h, heat) (K_m, momentum)
L	Obukhov length: $L = -u^{*3}/(kH^*)$
n	A generalized exponent
N	Brunt-Vaisala frequency: $N^2 = (g/\theta)\partial\theta/\partial z$
NBL	Neutral Boundary Layer
o	Subscript refers to surface value
p	Atmospheric pressure
P/G/T	Pasquill/Gifford/Turner
P_o	Standard atmospheric pressure (1000 mb)
Q	Source strength (rate of release)
r	Correlation coefficient
R_e	Reynolds number
R_i	Gradient Richardson number

R_b	Bulk Richardson number
SBL	Stable Boundary Layer
t	Time
T	Absolute temperature
T_e	Eulerian time scale for ABL turbulence
T_l	Lagrangian time scale for ABL turbulence
T_v	Virtual absolute temperature
u	Wind speed in the downwind direction (usually defined by mean wind direction at source height).
$\overline{u}$	Mean wind speed in the downwind direction
u_g	Geostrophic wind speed - the forcing formation for the wind
u^*	Friction velocity: $u^{*2} = -\overline{w'u'} = \tau_0$
U	Average $\overline{u}$ through mixed depth
v	Wind speed in the lateral direction
w	Wind speed in the vertical direction
w^*	Convective scale velocity: $w^* = (H^* z_i)^{1/3}$
x	Distance downwind of source
y	Lateral distance from plume centerline
z	Height above surface
z_i	Mixing depth in CBL, usually capped by an inversion
Z_o	Roughness length
Z_s	Source height
β	Average slope of a site
θ	Potential temperature, $T(p_o/p)^{2/7}$
$\Delta\theta$	Difference in θ
σ_a	Standard deviation of azimuth angle of wind

σ_e	Standard deviation of elevation angle of wind
σ_u, σ_v, σ_w	Standard deviation of wind components u, v, or w
σ_v, σ_y, σ_z	Standard deviation of concentration in lateral (y) and vertical (z) direction
ρ	Density of air or fluid
τ	Horizontal kinematic stress, time scale
χ	Concentration
ν	Kinematic viscosity

ACKNOWLEDGEMENTS

The authors wish to express their deep appreciation to John S. Irwin of the Meteorology and Assessment Division for his penetrating review and constructive suggestions. We also express our profound gratitude to Barbara Hinton and Brian Eder for their Herculean labors in typing this manuscript and preparing the figures, respectively. Finally, since no task can be finished without supervisory support and guidance, we wish to take this opportunity to thank John F. Clarke and Francis A. Schiermeier for their patience and support, especially during the final phase of delivery of this document.

Further we wish to thank Evelyn Poole-Kober for her work in obtaining the references and for proofreading the manuscript.

SECTION 1

INTRODUCTION, PURPOSE AND SCOPE

The basis of most applied diffusion modeling today, excluding buoyancy effects, was developed in the late 1950s and early 1960s. These Gaussian-empirical models were built on the results of a few field experiments that were quite limited in scope and that had problems with the tracers employed and instrumental accuracy, although they represented the state of the art at the time. Those results have been extrapolated to distances 15 to 125 times as large as the range of measurements and to more extreme stabilities; furthermore, they have been applied to much rougher or much more hilly sites and to much higher sources than used in the original experiments. Some of these extrapolations were guided by statistical theory, but most were pencil-and-straightedge or freehand extrapolations. On this basis, hundreds of multi-million dollar decisions were made and continue to be made.

This extrapolative situation, based on such limited experimental data, did not come about for lack of progress in diffusion science since 1960. In Section 3, over 20 major field experiments and numerous laboratory experiment are reviewed that were made after the two experiments that support most applied diffusion modeling today. In Section 2, large advances in knowledge about atmospheric boundary layers that have been made since 1970 are reviewed; these advances provide much support for better diffusion modeling. In Section 5, it is seen that classical diffusion theories have been further extended and that many new theoretical techniques with considerable validation have been developed as a result of availability of faster and cheaper computers. Hundreds of new papers on diffusion experiments, theory, and model validation appear every year in conference volumes or in journals. Thus, much progress has occurred since 1960 at the research level.

There is a perception in the scientific community that the best of the newest work is not being assimilated into the regulatory process in a rapid enough manner. A counterview is that to effect the transfer of new technology into the regulatory process requires more than a demonstration of a new idea with a single data set; a proof of superiority over existing techniques in wide variety of cases is also necessary. Historically, only a small fraction of the resources for research has been committed for this transfer.

The purpose of this paper is to provide an evaluation of what has been accomplished by past diffusion research and what most needs to be done. The scope is purposely limited to "basic" atmospheric boundary-layer diffusion, leaving out the complications introduced by source buoyancy, plume deposition, chemical reactivity, and complex flows due to structures, large terrain features, and land-water boundaries. This limit is imposed because basic boundary layer diffusion, in itself, involves complex phenomena and a large body of literature (the few hundred references used in this report represent only the most classic or most recent research advances). Also, a good understanding of basic diffusion must be the starting point for advances in modeling the more complex diffusion problems. Each of these problems can best be addressed separately, by specialists, as some already have. Most basic and specialized areas in diffusion research were reviewed in Atmospheric Science and Power Production (Randerson, 1984), and complex terrain models were recently assessed by Schiermeier (1984).

SECTION 2

ATMOSPHERIC BOUNDARY LAYERS

What have we learned about their dispersive properties?

The atmospheric boundary layer (ABL) refers to that part of the atmosphere directly affected by turbulent mixing near the surface. Heat, water vapor, pollutants, and momentum mix through this layer much more readily than through the air above it, which is only sporadically turbulent (in clouds and in "clear air turbulence"). This layer of continuous turbulent mixing, the ABL, is virtually always present, although its depth ranges widely from day to night, season to season, and place to place (it is absent only on windless nights in topographic "bowls" of cold, dense air). It is also the part of the atmosphere in which nearly all human activity takes place, including release of pollutants. Fortunately for us, turbulence is effective at dispersing and diluting pollutants; without it, molecular diffusion would diffuse a point-source plume to a height and width of only 4 m in 24 hr (edges defined by 50% of centerline concentration). Unfortunately for those people trying to predict turbulent diffusion, the rate at which it proceeds varies quite widely, depending on wind, atmospheric stability, the height of the ABL, the height of release of pollutants and surface characteristics in a rather complex way.

The ABL may be viewed as the place where the large scale pressure systems constituting "weather" interface with the planet's surface. The geostropic wind, u_g, is a measure of forcing by the pressure field, which changes as the large migratory systems ("highs" and "lows") move across the land. The surface fluxes of heat, moisture, momentum, and pollutants into or out of the ABL are all strongly influenced by surface cover and local land-use (a pasture has very different texture and thermal properties from those of an industrial area, for example). It is in the ABL that the influences of large scale forcing and local surface fluxes are recounciled; this process accounts for the great variability in wind, temperature, and diffusion rates which we observe.

The discussion that follows begins with definitions and explanations of all meteorological variables important to diffusion in ABLs. Then the typical diurnal cycle of meteoroloigcal change and some of its effects on diffusion is described, with particular attention to transitional periods. Finally, current understanding of ABLs, their turbulence structure and their diffusive abilities are described. Because stability has the greatest effect on diffusion rates and because there are basic structural differences in the turbulence of stable boundary layers (SBLs), neutral boundary layers (NBLs), and unstable or "convective" boundary layers (CBLs), it is convenient to consider these separately.

IMPORTANT METEOROLOGICAL VARIABLES AND SCALES AFFECTING DIFFUSION

Regardless of the stability, the same basic collection of meteorological variables are of prime concern for diffusion modeling. They will be introduced in this section. Overbars indicate a time average, and primes indicate deviations from the time average. Subscript "o" refers to surface values.

There are several ways to measure stability. One measure is the static stability, which in an incompressible fluid is $-\partial\rho/\partial z$, where ρ is the fluid density and z is height above the surface. This can be used as a stability measure in air, but the more easily measured quantity is $\partial T/\partial z$, where T is absolute temperature (air is nearly an ideal gas, so at constant pressure, $\rho \propto 1/T$). Because air is compressible, a better stability measure is $\partial\theta/\partial z$, where θ is the "potential" temperature defined by $\theta = T(p_0/p)^{2/7}$ (p is ambient pressure, and p_0 is a standard, sea level value, usually 1000 mb). This quantity compensates for heating and cooling due to pressure changes, so that if air rises or falls without radiating or absorbing heat its θ is constant, i.e., θ is a conservative property (this type of process is called "adiabatic"). If $\partial\theta/\partial z = 0$, the stability is "neutral", meaning that if air is displaced vertically, it remains at the same θ, T, and ρ as ambient air on that constant pressure surface (p is primarily a function of z); thus, it develops no buoyancy. However, if $\partial\theta/\partial z > 0$, upwardly displaced air will find itself surrounded by higher θ, higher T, and lower density air; therefore it will be heavier than the ambient air. This displaced air acquires buoyancy, or negative buoyancy, which tends to return it to its original height; hence, the air is "stable". If $\partial\theta/\partial z < 0$, the opposite happens, and displaced air acquires buoyancy acting in the same direction as the displacement. This displaced air tends to keep rising or falling, even accelerating, and the air is called "unstable". Turbulence develops spontaneously in unstable air; this is called "convective" turbulence.

There is a simple relationship between the vertical gradients of θ and T, derivable from the definition of θ, the ideal gas equation of state, and the hydrostatic equation, $\partial p/\partial z = - \rho g$, where g is gravitational acceleration (this equation follows from the fact that static pressure is just the weight per unit area of the air overhead). For the earth's atmosphere, the result is

$$\partial\theta/\partial z = (\partial T/\partial z + 0.01°\mathrm{C/m})\,(p_0/p)^{2/7} \tag{2.1}$$

Naturally, the larger the $\partial\theta/\partial z$, the more stable the air is; $0 < \partial\theta/\partial z < 0.01°\mathrm{C/m}$ is considered slightly stable; $\partial\theta/\partial z > 0.01°\mathrm{C/m}$ is usually considered moderately stable, unless $\partial\theta/\partial z > 0.03°\mathrm{C/m}$, which is roughly the dividing line for "very" stable. However, static stability is not the only consideration. A vertical gradient of mean wind speed, $\partial\overline{u}/\partial z$, tends to encourage overturning and the development of turbulence. If equal amounts of air from two layers mix, one layer with a mean wind speed $\overline{u} = u_1$ and the other with $\overline{u} = u_2$, the conservation of momentum principle gives $\overline{u} = (u_1 + u_2)/2$ for the mixture. However, the kinetic energy (KE) of mean motion, $\overline{u}^2/2$, is less than the original amount, $(u_1{}^2/2 + u_2{}^2/2)/2$, by the amount $(u_2 - u_1)^2/4$. This surplus KE feeds turbulent motion, i.e., it becomes turbulent KE. This source of turbulent KE is present whenever $\partial\overline{u}/\partial z$ is not zero, regardless of sign. These considerations lead to a very useful stability index, the gradient Richardson number:

$$Ri = (g/\theta)(\partial\theta/\partial z)/(\partial u/\partial z)^2 \tag{2.2}$$

This number is dimensionless; it compares the potential energy increase needed to mix stable air with the surplus KE that would be available after mixing due to $\partial\overline{u}/\partial z$. When $Ri > 1$, turbulence is extinguished rapidly, as there is not enough surplus KE to drive air mixing "uphill" against the positive $\partial\theta/\partial z$. At $Ri = 0.5$ there is an approximate energy balance, and at $Ri < 0.25$, turbulence develops spontaneously from the slightest perturbation, even if $\partial\theta/\partial z$ is positive. Thus, Ri is a physically meaningful stability index; it also highly correlates with diffusion rates in certain cases, which are discussed in Section 4.

Wind speed and its fluctuations are very important quantities in diffusion modeling because they have four effects: transport, axial dilution (for continuous releases), shearing of the plume or puff, and turbulence-induced growth and dilution. The usual meteorological convention is to define x as the longitudinal horizontal spacial coordinate (extending in the direction of the mean wind), y as the lateral horizontal coordinate (perpendicular to the mean wind), and z as height above the surface. For a point source, we define $x = 0$ and $y = 0$ at the source. The corresponding components of velocity are u, v, and w, respectively. The direction of x is generally defined by the mean wind direction at the source height, so that $\overline{v} = 0$ at that height. However, if the wind direction turns with height, $\partial\overline{v}/\partial z \neq 0$; this tends to increase the plume width with distance, as the plume top is transported in one direction, and the plume bottom is transported in another. Turbulent velocities are defined by deviations from the time average, e.g., $v' = v - \overline{v}$. Averaged over the same time span, or "sampling time", we must have $\overline{u'}$, $\overline{v'}$, and $\overline{w'} = 0$. The mean motion in the atmosphere closely parallels the ground, so over flat ground we usually can assume that $\overline{w} \approx 0$ also. Turbulence velocity standard deviations are defined by $\sigma_v = (\overline{v'v'})^{1/2}$, etc. These are very important quantities in diffusion modeling, as the classical statistical theory prediction (Taylor, 1921) for diffusion close to the source is well proven: $\sigma_x = \sigma_u t$, $\sigma_y = \sigma_y t$, and $\sigma_z = \sigma_w t$, where t is the time since release and σ_x, σ_y, and σ_z are spacial standard deviations of concentration distribution from point source.

Except for cold air drainage and warm air rising on slopes, the main force driving the ABL is the horizontal pressure gradient, $\partial p/\partial x$ and $\partial p/\partial y$. The ABL is also affected by the turning of the earth; relative to fixed coordinates on the earth's surface, this turning causes an apparent acceleration to the right (in the northern hemisphere) of any horizontal motion. This "Coriolis" acceleration is equal to the horizontal speed times the Coriolis parameter, f, which is twice the angular rotation rate of the earth ($2\pi/24$ hr) times the sine of the latitude; $f^{-1} \simeq 10^4$ s at mid-latitudes. Above the ABL, where there are no appreciable surface friction effects, the pressure gradient acceleration approximately balances the Coriolis acceleration. This balance defines the geostropic wind speed, u_g, which is usually closely approximated by the real wind at those heights (e.g., if $\partial p/\partial x = 0$, $u_g f = -\rho^{-1}\partial p/\partial y$). Thus, u_g and f can be regarded as external parameters important to the ABL. The surface kinematic stress, τ_0, which is the same as "friction", defines an important internal parameter, the friction velocity: $u* = (\tau_0)^{1/2}$. Surface friction is equal to downward vertical turbulent flux of horizontal momentum into the ground, which gives an alternative definition of u*: $u*^2 = -\overline{w'u'}$ measured near the surface. One other parameter is also important in describing surface friction effects, the roughness length, z_0. It can be determined experimentally from wind speed profiles near the ground, and it scales roughly as about 1/10 the height of most "drag elements" on the surface (trees, grass, crops, buildings, etc.); z_0 can be crudely estimated by inspection of a site (Weiringa, 1980). The velocity scales u_g and u^* are roughly proportional, but u_g/u^* is influenced by z_0 and by the stability.

Turbulent velocities and diffusion near the surface, especially their vertical components, are strongly correlated with u^* and the flux of buoyancy due to heating or cooling of the surface. This flux by $(g/T_v)\overline{w'T'_v} = H^*$, where T_v is the virtual absolute temperature, which includes the effect of water vapor on air density ($T_v = T/(1 - 0.6q)$, where q is the specific humidity). Except over very moist surfaces, the humidity effect is small and H^* may be determined by using only the "dry" temperature, T, in place of T_v. H^* represents the turbulent flux of positive buoyant acceleration caused by heating of the air (or vaporizing of water) next to the ground; it is proportional to the upward sensible heat flux plus about 1/7 of the latent heat flux (Briggs, 1985a). When the surface is colder than the surface air, the heat flux is downward, and H^* is negative. H^* is also equal to the rate per unit mass at which turbulent energy is created by buoyancy; it is the rate of potential energy release in vertical, turbulent motions. When H^* is positive, $\partial\theta/\partial z$ becomes negative due to stronger heating near the surface, and convective turbulence is produced. When H^* is negative, $\partial\theta/\partial z$ becomes positive due to stronger cooling near the surface, and H^* represents a rate of loss of turbulent energy. When turbulence is produced by wind shear, it is called mechanical. The rate of mechanical production of turbulent energy is $\tau\partial u/\overline{\partial z}$, so $u^{*2} = \tau_0$ also relates to turbulence production. Both u* and H* can be measured using fast-response measurements of u', w', and T' (or T_v').

Together, u* and H* form an important length scale, the Obukhov length (Monin, 1959), $L = -u^{*3}/(kH^*)$, where $k \simeq 0.4$, the von Karman constant (Hogstrom, 1985); this length is defined negative when H* is positive (unstable conditions). Near the surface, it is well established that $\partial\overline{u}/\partial z \simeq k^{-1}u^*/z$, so the ratio of buoyant to mechanical production of turbulent energy is $H^*/(u^{*2}\partial\overline{u}/\partial z = -z/L$. Thus, $z = |L|$ is a dividing line between dominance of buoyancy effects (above $|L|$) and mechanical effects (below $|L|$). According to surface-layer similarity theory, discussed here in Section 5, $\partial\overline{u}/\partial z = u^*/(kz)$ times a function of z/L and $(g/\theta)\partial\theta/\partial z = H^*/(ku^*z)$ times another function of z/L; these hypotheses have been well supported by field measurements (e.g., Businger et al., 1971). It follows that in the surface layer (the lowest 10% or so of the ABL), R_i is a function of z/L. It is also possible to relate either Ri or L to the Bulk Richardson number, $R_b \propto \Delta\Theta/\overline{u}^2$, by using surface-layer similarity equations for $\overline{u}$ and θ (Irwin and Binkowski, 1981). Both L and Ri have been shown to be highly correlated with vertical diffusion from surface sources (e.g., Weber et al., 1977). This is consistent with the fact that vertical turbulence quantities, e.g., σ_w/u^*, have been shown to be functions of z/L.

The wind speed gradient in the surface layer, $kz\partial u/\overline{\partial z} = u^*\phi_m$, where ϕ_m is a function of z/L, can be integrated to give

$$\overline{u} = (u^*/k)[\ln(z/z_0) - \phi_m], \tag{2.3}$$

where ϕ_m is a function of z/L, given by the integral of $(1 - \phi_m)d\ln(z/L)$. This equation provides a way to calculate $\overline{u}$ for an elevated release using $\overline{u}$ measured closer to the surface, if we can measure or estimate L.

The remaining parameter that is of great importance in describing the diffusion potential is the height of turbulence mixing in the ABL. For CBLs this is usually designated as z_i (most CBLs are capped by a temperature inversion, i.e., positive $\partial T/\partial z$, hence the subscript "i"). For SBLs this is usually designated as h, which will be used here for NBLs also. First of all, it is important to know whether an elevated source is within or above the mixing layer, as there is normally no turbulent diffusion above this layer. Except in clouds and in occasional patches of "clear air turbulence", the atmosphere above the ABL is always stable and non-turbulent, regardless of stability swings at the surface (the upper air is not subject to intense heating and cooling like the air next to the surface). Especially on clear nights with low u_g, when h can be quite small, elevated and even medium-level sources may be above h. Second, turbulence intensity and structure is a function of z/z_i or z/h; the intensity falls off rapidly with height near z_i or h. Third, in the CBL it has been found that the size of the largest turbulent eddies approximates z_i and that σ_u and σ_v are proportional to the scale velocity $w^* = (H^* z_i)^{1/3}$ throughout the mixing layer.

In summary, in the order of discussion, the meteorological variables and scales most relevant to diffusion are $\partial T/\partial z$, $\partial \theta/\partial z$, Ri, $\overline{u}$, $\overline{v}$, σ_u, σ_v, σ_w, f, u_g, u^*, z_0, H^*, L, and z_i or h.

THE DIURNAL CYCLE AND TRANSITIONAL PERIODS

Before discussing neutral, convective, and stable boundary layers considered separately, it is worthwhile describing what is known about the normal diurnal sequence of stability changes in the ABL. This description excludes overcast, windy conditions, which tend towards neutral regardless of the time of day. In the course of a typical day, the ABL goes through a wide range of structure, being stable at night and convective during most of the daylight hours. During the transitional periods around sunrise and sunset, rapid changes occur. Rapid changes also occur during thunderstorms, at the onset of sea breezes, and with frontal passages. Because these latter events are accomplished by brisk wind speeds, they are not likely to be causes of serious pollution episodes. The steady state assumptions customarily made in diffusion modeling can be more or less justified for 1-hr averaging during the major phases of the 24-hr cycle, but have little validity during the transitional periods, as will be seen. These transitional periods have received less study than the major ABL states, but they can be described at least qualitatively.

A good time to begin this description is the hour before sunrise, because this is ordinarily a time of steady state. Radiative cooling has created stable stratification in the lowest 300 to 1000 m, and there may be a remnant of the inversion above the previous day's maximum z_i (radiative cooling can intensify the inversion if there was much haze below that z_i). There is stronger stability near the surface and a shallow mixed layer to h produced either by (1) wind shear-generated turbulence, especially when u_g is large, (2) slope flows generated by cooling of the surface air and gravity, especially when u_g is small, or (3) convective mixing produced by more intense radiational cooling at the top of a fog extending to depth h, especially when u_g is small in humid conditions on clear nights.

When sunlight first reaches the surface, at low angle, most of its energy goes at first into evaporating dewfall or fog; this latent heat produces some increase in buoyancy, as water vapor is lighter than air, but only about 1/7 the buoyancy produced by the sensible heating of air. Consequently, there is not much convective activity until most of the liquid water or frost has evaporated. This may take no time at all in a desert, about an hour for typical dewfall or frost in a humid climate, or half a day for a deep fog. After this evaporation is accomplished, a much larger fraction of solar energy ends up as sensible heat in the air next to the surface. The air becomes lighter and rises; small thermals converge to produce larger ones, and these impinge against the top of the mixed layer, overshoot a little into the stable air above, mix with some of it, and bring it back down as the mixture subsides and feeds into downdrafts. This is the beginning of the development of the daytime CBL. The thermals cause large undulations in z_i, especially in the first hours of CBL development. If there happens to be an elevated layer containing pollutants, as occurs when a buoyant plume rises and stratifies in non-turbulent air at night, the pollutants, too, are entrained by invading thermals and are mixed down to the ground in the "inversion breakup fumigation" process. The depth z_i continues to grow until early to midafternoon when solar heating and H* begin to diminish. Sometimes during the day z_i appears to "jump", as thermals can traverse a near-neutral layer aloft quite easily once they reach nearly the same potential temperature as the neutral layer. The average θ is nearly uniform through the bulk of the mixed layer, and it increases steadily through the day as long as H* is positive (until about 1 h before sunset). The heating rate diminishes through midday and afternoon as H* decreases and the depth being heated, z_i, maximizes. The negative $\partial\theta/\partial z$ layer near the ground reaches maximum in intensity near noon, when H* is at a maximum. As H* decreases in midafternoon, this negative $\partial\theta/\partial z$ decreases, so the surface temperature may begin to decrease even though H* is still positive and θ is still slowly increasing above 0.1 z_i.

Exactly what happens in the CBL in the very late afternoon is still an area of speculation. In clear conditions, H* goes to zero about an hour before sunset, as outgoing thermal radiation from the warmed surface nearly balances incoming solar radiation. The CBL may begin to collapse before H* goes to zero as the thermals become weak. Convective downdrafts outside the thermals have a slight stable stratification due to positive $\partial\theta/\partial t = -w\partial\theta/\partial z$. When new thermals are very weak, they cannot penetrate through this stable stratification all the way to the previous z_i, so the mixing depth descends. It is not known how rapidly this occurs. The continuation of positive moisture flux from the surface during z_i collapse may produce a negative humidity gradient, which leads to more rapid radiative cooling at lower heights, and further stable stratification that will hasten the demise of turbulence aloft. In dry conditions, the CBL collapse process may go slower. Without cooling due to radiative flux divergence (due mostly to humidity), the stratification remains close to neutral to z_i

and the turbulence gradually "runs down", with a time scale ~ z_i/σ_w, on the order of 1 hr in the late day.

While what remains of the CBL aloft is running down or collapsing due to stratification developing from radiation flux divergence, the SBL begins development before sunset, as soon as H* turns from positive to negative. As the sun sinks, incoming solar radiation falls to zero, while outgoing long-wave radiation from the surface continues all night. Because the surface emits almost like a black body, this radiation flux is $\propto T_s^4$, where T_s is the absolute surface temperature. In spite of nocturnal cooling, absolute temperature does not change much, so the radiation balance at the surface is rather constant on clear nights (if a cloud deck moves over the site, it radiates back like a black body at slightly smaller T^4, so the outgoing surface radiation is almost balanced and the net radiation is reduced drastically). The net radiative cooling at the surface is divided between cooling the ground slab, condensing or freezing water in any dew, fog, or frost, and cooling the air next to the ground. If it is very humid, as after a summer shower, fog or dew can form right away, but usually this does not occur until later at night as the surface temperature approaches the dew point; of course, this does not happen at all in dry conditions. The soil heat flux to the surface is fairly constant through the night, but the surface cooling rate decreases as the heat conducts out of an increasingly deeper ground slab (Garratt and Brost, 1981).

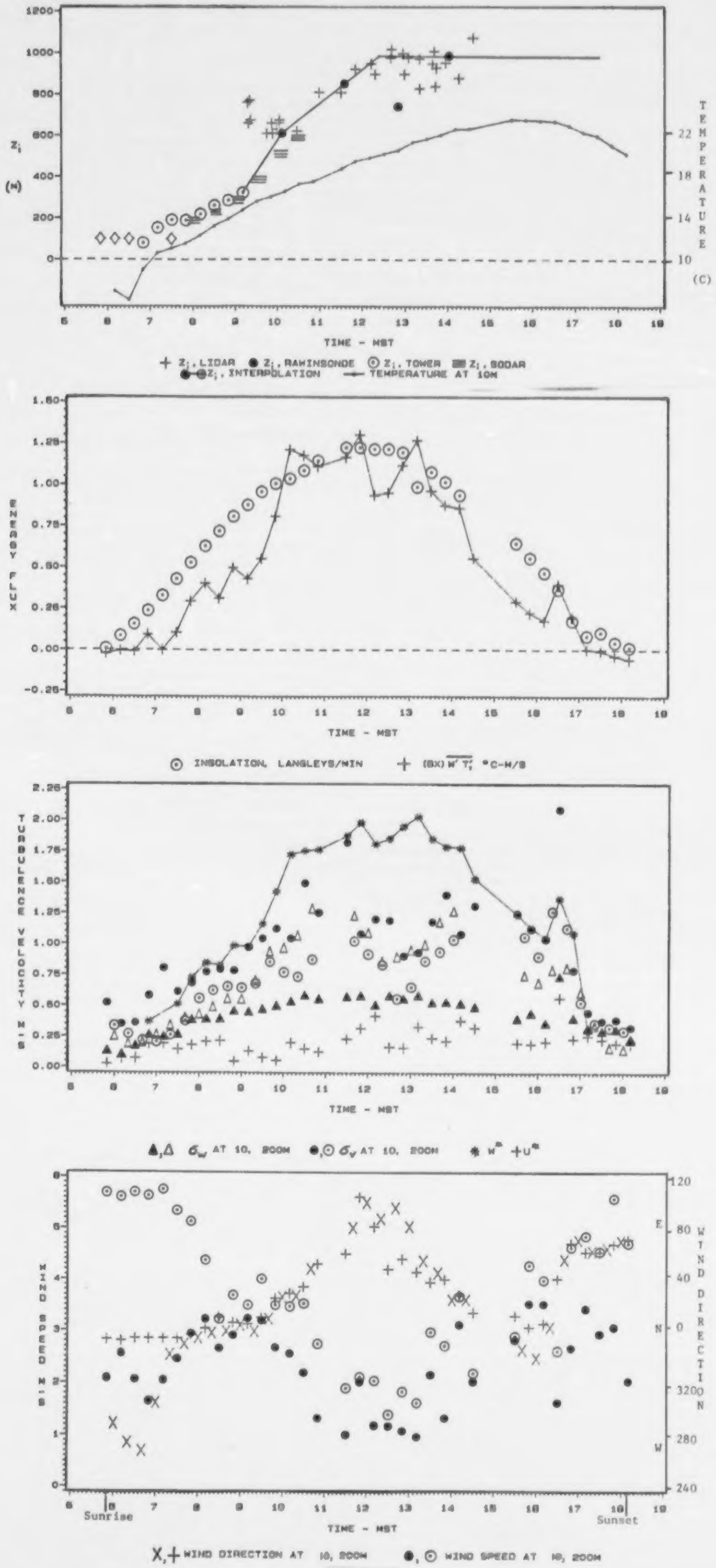
Surface parameters as well as SBL structure evolve rapidly in the first few hours after transition. This is evident from the results obtained with second-order closure ABL models (Wyngaard, 1975; Brost and Wyngaard, 1978), which appear to give good simulations of observations from the 1973 Minnesota experiment (Rao and Snodgrass, 1978; Wyngaard, 1975). The heat flux becomes increasingly negative up to a point, about 1 hr after transition, and then gradually diminishes in magnitude in spite of the relative constancy of the net radiation flux. At the same time, u*, L, and h rapidly reduce from their values at t = 0 (transition), but they may approach a rough steady state in 3 or 4 h, if not interrupted by a turbulent burst (a sporadic episode of turbulent mixing due to changes in the structure of Ri). This probably happens because the increasing stability near the surface restricts turbulent motions, so that u* drops, and as turbulence at the surface diminishes, the heat transfer from surface to air is restricted, which causes H* to drop. (Heat must first pass through a "laminar sublayer" whose thickness is proportional to ν/u^*; this is the same insulating layer that our bodies lose at higher wind speeds, hence, the "wind chill" factor.) The bulk Richardson number for the whole layer gives some insight into the behavior of h. Wind shear can support turbulence working against stable stratification only as long as $R_b = (g/\theta)(\theta_h - \theta_o)h/\overline{u}_h^2$ is less than 0.3 or so. The surface air cools more rapidly than air higher up, so if $\overline{u}_h$ remains constant, the increasing $(\theta_h - \theta_o)$, which increases the "heaviness" of the surface air, pulls h downward. Of course, $\overline{u}_h$ can also change as h moves downward, or as momentum is transferred in gravity waves or due to changes in driving forces like $u_g f$. Thus, h can behave rather irregularly through a night (e.g., see Fig. 2). In some of the Minnesota runs, u_g dropped during the night and turbulence became immeasurable after about midnight. In such cases, we might as well forget about turbulent diffusion until sunrise, and just try to determine in which direction a plume of pollutants is heading and whether it is spreading due to horizontal eddies or meandering. In this situation, there is no substitute for wind speed and direction measured near the point of release.

Several efforts have been made to mathematically simulate the complete diurnal behavior of the ABL, e.g., Yamada and Mellor (1975) and Binkowski (1983). Both of these simulations show a rapid decay of CBL turbulence in the late afternoon, and predict a maximum in the negative surface heat flux in the SBL about 2 hr after transition to negative values.

The temporal evolution of a fairly typical CBL, as shown by actual measurements, is plotted in Fig. 1. The measurements are from the Boulder Atmospheric Observatory of NOAA's Wave Propagation Laboratory during the CONDORS experiment (Moninger et al., 1983). The day, 13 Sept. 1983, was cloudless and very convective; $\overline{u}$ dropped as low as 2 m/s during the midday, giving $\overline{u}/w^* \simeq 1$ and $|z_i/l|$ exceeding 100. Fig. 1a shows the development of z_i and T at 10 m. For about an hour after sunrise there is evidence of continuation of a drainage wind from the west up to a height of about 100 m (this flow is peculiar at this site and probably originates from the east slope of the Rocky Mountains, which begin 25 km to the west). This flow breaks up soon after the heat flux turns positive (Fig. 1b) and z_i climbs more or less steadily until noon, at which time it seems to reach its limit of $\simeq$ 1000 m for that day (z_i sometimes leveled off for several hours during

Figure 1

BAO



Z_i (M)
TEMPERATURE (C)
TIME - MST
+ Z_i, LIDAR ● Z_i, RAWINSONDE ⊙ Z_i, TOWER ▤ Z_i, SODAR
●—● Z_i, INTERPOLATION —+— TEMPERATURE AT 10M
ENERGY FLUX
TIME - MST
⊙ INSOLATION, LANGLEYS/MIN + (8X) $\overline{w'T_i'}$ °C-M/S
TURBULENCE VELOCITY M/S
TIME - MST
▲,△ σ_w AT 10, 200M ●,⊙ σ_v AT 10, 200M * w^* + u^*
WIND SPEED M/S
WIND DIRECTION
E
N
W
Sunrise
Sunset
TIME - MST
X,+ WIND DIRECTION AT 10, 200M ●, ⊙ WIND SPEED AT 10, 200M

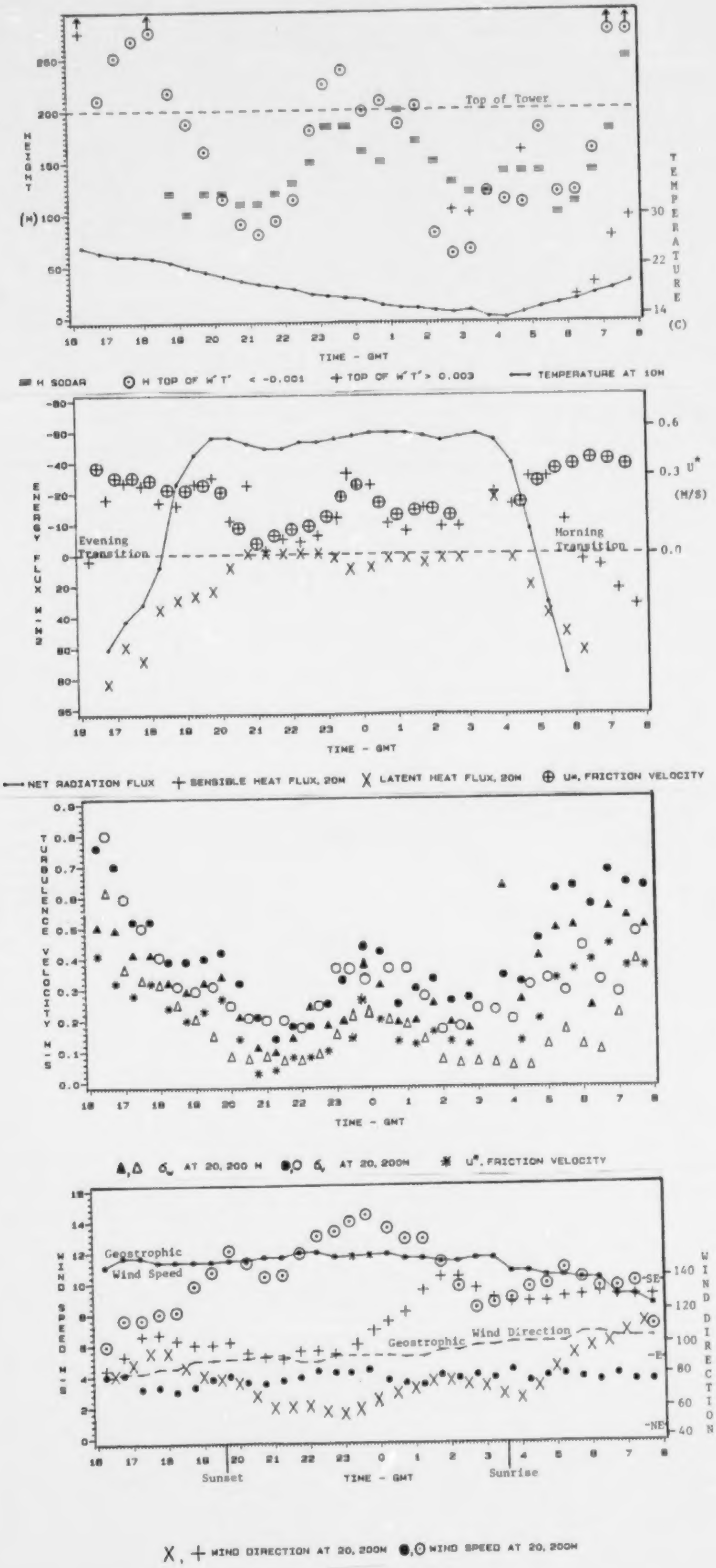
this experiment, and on some days it appeared to grow in stair-step fashion). The scatter in the lidar measurements, which were made at various azimuths and shot elevations, give an idea of the magnitude of undulations in local z_i, perhaps ± 100 m. In Fig. 1b we see that the sensible heat flux ($\propto \overline{w'T'}$) is very much controlled by the insolation, although it lags behind insolation in the first hours of daylight; this lag may be due to both outgoing thermal radiation from the surface to the clear sky and channeling of solar energy into evaporation of wetness in the ground cover (predominantly deep grass).

Figure 1c shows turbulence velocities at 10 and 200 m together with w^* and u^*. Although there is much scatter in the σ_v values and the σ_w values at 200 m, after z_i passes through 200 m at about 8 AM, they appear to be essentially controlled by w^*. The 20-min average u^* values are much smaller and fluctuate widely. Note that w^* is nearly constant from 10:10 to 14:10. In contrast, σ_v and σ_w at 200 m appear to go through an oscillation, peaking near 11:00 and 14:00; this may be related to the several-hour long cycles in some CBL variables reported by Kaimal et al. (1976) from the Minnesota experiment. We note that σ_w at 10 m is relatively steady, with only a slow rise and fall at the beginning and end of the day; it behaves like the free-convective prediction, $\sigma_w \propto (H^*z)^{1/3}$, and seems to experience very little perturbation from the passing large eddies. (At noon on this day a typical downdraft 1.5 z_i wide and moving at 2 m/s would take 12.5 min to pass the tower; therefore, it is not surprising to see much scatter in 20-min averages.) In Fig. 1d we see how well the 10- and 200-m wind directions track each other through rather large changes after 8 AM, so the wind direction shear is negligible in this layer during CBL hours. Before CBL development, the wind was from the north at $\sim$5.6 m/s at 200 m and from the west at $\sim$2.1 m/s at 10 m, in the presumed drainage flow. The wind directions become easterly (from the east) at midday, probably due to forcing from the upslope flow on the east slope of the Rockies to the west, and phenomenon absent at sites far from mountains. The wind speed at 200 m is quite steady until z_i approaches that height at 07:20; then it drops and fluctuates widely during the most convective hours until it regains some of its steadiness at $\overline{w'T'}$ and w^* become negligible about 1 hr before sunset.

The temporal evolution of a SBL at a flat site (Cabauw, the Netherlands), as shown by measurements, is plotted in Fig. 2. The night, 30-31 May 1978, was clear and the large scale driving force was fairly constant, as is shown by the geostrophic wind speed and direction in Fig. 2d. In spite of this we see in Fig. 2a large oscillations in h, as determined by the top of the layer with significant downward heat flux. Note that the acoustic sounder or "sodar" representation of h is in general agreement, but does not fluctuate as widely as h determined by $\overline{w'T'} < 0.001$°C-m/s (this determination was made by extrapolation in the range 200-270 m). The temperature at 10 m falls almost steadily until 45 min after sunrise, and then rises rapidly and steadily. Various energy fluxes near the surface are plotted in Fig. 2b. The net (incoming minus outgoing) radiative flux crosses to negative about 1 1/4 hr before sunset and then becomes practically constant through this clear night; it crosses to positive about 1 1/4 hr after sunrise, rapidly increasing. The tranpirative response of the grass and trees may account for the continued upward flux of water vapor (positive

CABAUW

Figure 2



latent heat flux) to 1 hr after sunset and its quick reversal back to upward flux minutes after sunrise. The early transition from positive to negative heat flux, about 3 hr before sundown, is surprising to us (the authors); perhaps the grass surface is cooling due to transpiration more rapidly than it absorbs solar heat (late afternoons in summer, grass often feels cool to bare feet). The delay in positive sensible heat flux until 2 1/2 hr after sunrise mirrors the evening transition behavior; this could be due to wind-induced evaporation of dew. During the SBL hours, Fig. 2b shows a striking correspondence between sensible heat flux and u*, supporting the constant $T^* = \overline{w'T'}/u^*$ approximation suggested by van Ulden and Holtslag (1983).

In Fig. 2c we see that variations through the night in turbulent velocities also strongly follow variations in u*, especially the velocity measurements at z = 20 m. However, σ_w at 200 m tends to not respond to u* and remains at a near constant low value during periods of $h < 200$ m (roughly 1900-2300 and 0200-0700 GMT). These small σ_w values may reflect wave motion rather than low level vertical turbulence in the air above h. Values of σ_v at 200 m are 2 to 4 times larger than σ_w during these periods and suggest a persistence of horizontal eddy intensity after h drops below 200 m. Note that from 1800 to 0430 the rises and falls in h correspond well to variations in σ_v, σ_w, and u* measured at 20 m, with the exception of a disturbance at 0345 (u* could not be meaningfully determined for this 30-min period, because the measured $\overline{u'w'}$ was positive, the "wrong" sign). After sunrise, a marked rise in u* seems to precede the measured point of transition to positive buoyancy flux at about 0550, indicating perhaps that the stability near the surface is weakening (T at

10 m rises with u*); less stability causes less suppression of vertical turbulence, which tends to increase surface friction. Immediately after transition to positive heat flux at 0600 (Fig. 2b) we see the rapid development of a CBL (Fig. 2a); in its early development, it is capped by a thick layer of negative heat flux.

In Fig. 2d we see relatively small differences between 20- and 200-m values of wind speed and wind direction just before evening transition and increasingly independent behavior afterward. This characterizes a principle difference between CBLs and SBLs. As h drops below 200 m, $\overline{u}_{200}$ approaches and oscillates around u_g, the geostrophic wind speed. (Some of these oscillations might be due to variations in the local u_g or pressure field; the measured values represent u_g on a large scale.) In constrast, $\overline{u}_{20}$ changes little during this night and is seemingly indifferent to changes in other variables. After evening transition the wind direction at 20 m begins turning to the left of the geostrophic wind, which means that it turns toward low pressure; this is normal behavior near the surface in a SBL. This wind direction difference begins to diminish in the morning at the same time that T_{10} and u* begin to rise, indicating increased coupling with upper flow at this time. The wind direction at 200 m approaches the large scale geostrophic wind closely until 2300, then turns to the right nearly 50°, an event for which we have no explanation.

Figures 1 and 2 illustrate dramatically the many complex changes that occur in ABLs around the times of morning and evening transition. They are also illustrative of many of the understood features of quasi-steady state

CBLs and SBLs, so will be useful references for the CBL and SBL subsections to follow. In addition to the understood features, measurement records like these often reveal events we cannot readily explain, such as the brief, marked increases in magnitudes of turbulence variables at 1630 MST in Fig. 1 and at 0345 GMT in Fig. 2.

NEUTRAL BOUNDARY LAYERS

A neutral boundary layer is one in which buoyancy forces are negligible. Thus, $\partial\theta/\partial z = 0$ throughout a NBL and temperature drops 1°C per 100 m increase in elevation. For this condition to hold near the surface, H^* must be zero or very small because at small z/L, $\partial\theta/\partial z = (\theta/g)H^*/(ku^*z)$. Also, the production (or dissipation) of turbulent energy by buoyant forces must be small compared to the production by mechanical forces (wind shear); this requires that $h < |L|$. This condition is met rather infrequently, which is perhaps why there has not been as much research on NBLs as on CBLs and SBLs. As $L \propto u^{*3}/H^*$, a combination of high winds and negligible heating or cooling of the ground is required to produce a neutral boundary layer. This latter requirement is most likely to be met during overcast conditions, when very little solar radiation reaches the ground and most upward long-wave radiation from the surface is balanced by downward radiation from the cloud base (clouds and most surfaces radiate approximately as black bodies). There are also moments of $H^* = 0$ just after sunrise and just before sunset in the normal diurnal cycle when it is not overcast, as H^* changes sign from positive in the day to negative at night. This condition is so transient that the ABL does not have time to adjust to any kind of neutral structure.

The NBL is driven by the pressure gradient; the magnitude of the acceleration in the direction of the gradient is u_gf, which follows from the definition of u_g. Balancing this when the NBL is in a steady state (no acceleration) are the Coriolis accelerations $\overline{u}f$, directed to the right of the wind direction in the northern hemisphere, and frictional drag. This drag is given by $\partial\tau/\partial z$, where τ is the horizontal stress; $\tau = u^{*2}$ at the surface and goes to zero at the top of the NBL. Near the surface, $\overline{u}$ is small, so the Coriolis acceleration is weak and $\partial\tau/\partial z$ nearly balances the pressure gradient acceleration, u_gf; higher up, the Coriolis acceleration is stronger and $\partial\tau/\partial z$ becomes weaker, so the wind turns more to the right and approaches the geostropic wind near $z = h$. The two velocity scales u_g and u^* are more or less proportional, but ratio is influenced by the surface Rossby number, $u^*/(fz_0)$. The ratio u_g/u^* ranges from about 20 over rough surfaces to about 50 over very smooth surfaces (water and snow) in neutral conditions (Counihan, 1975).

Turbulence velocities in the NBL are most closely related to u^*. The vertical velocity standard deviation, given by $\sigma_w = (\overline{w'w'})^{1/2}$, is about $1.3u^*$ near the surface (Hanna, 1981). The lateral (across the wind) and longitudinal (along the wind) components of velocity variance, σ_v and σ_u, respectively, also scale to u^*, but observations show some variability in the ratios σ_v/u^* and σ_u/u^* at different sites. This may be caused by surface inhomogeneities or topographic influences; σ_v/u^* ranges from about 1.3 to 2.6 and σ_u/u^* ranges from about 2.1 to 2.9 near the surface (Lumley and Panofsky, 1964). Observations of these quantities high up in the NBL are lacking, but Hanna (1981) suggested exponentially decreasing values using the scaling height u^*/f; this is as an approximate fit to a second-order closure numerical model of the ABL (Wyngaard et al., 1974). Hanna also suggested an explicit form for the Lagrangian time scale; this is roughly the time it takes diffusing particles to "forget" the turbulent velocity they experienced at release, and it is an important time scale in diffusion theory (especially statistical theories.)

The height of the NBL is of concern, since it is where turbulence velocities drop off and where the wind speed approaches the geostrophic value. It is usually assumed that $h \propto u^*/f$, but this can only be true if $\partial\theta/\partial z \simeq 0$ through a deeper layer, so that h is not limited by stable capping. In this case, Deardorff's (1972) large eddy numerical modeling indicates that $\overline{u} \simeq u_g$ at a height $0.2u^*/f$, but turbulence velocities fall off gradually, rather than abruptly. His results are in rough agreement with Hanna's (1981) approximations, $\sigma_v/u^* = \sigma_w/u^* = 1.3 \exp(-2zf/u^*)$, except that Deardorff's σ_v does not fall off significantly from its surface value until $z > 0.17u^*/f$. Typically, $u^* \simeq 0.5$ m/s with $\overline{u} \simeq 5$ m/s at $z = 10$ m, and $f \simeq 10^{-4}$ s^{-1} at midlatitudes, so $0.2u^*/f$ is $\simeq 1000$ m. Obviously, u^*/f cannot be the correct form of h near the equator, because there $f = 0$. NBLs can also be limited by stable capping at $z = z_i$, just like CBLs, especially when $z_i f/u^*$ is small. Krishna and Arya (1981) investigated this case with an eddy diffusivity model. They concluded that there is no significant entrainment at z_i when $z_i > 0.2u^*/f$, but there is significant entrainment when $z_i = 0.05u^*/f$, along with other changes in boundary layer structure. Yokoyama et al., (1977) analyzed aircraft turbulence measurements in five early-morning to mid-morning NBLs capped by an inversion, for which hf/u^*

ranged from 0.09 to 0.16 (h was defined by the zero intercept of measured turbulence energy dissipation rate profiles; $h \approx z_i$ within +150 m and -50 m). For these boundary layers, σ_w was observed to be proportional to $(1 - z/h)$, a linear dropoff, in contrast to the exponential dropoff predicted by NBL models with $z_i f/u^* = 0.45$ or larger.

CONVECTIVE BOUNDARY LAYERS

Our understanding of CBLs, which prevail during the daytime, has greatly improved since convective scaling with z_i and $w^* = (H^* z_i)^{1/3}$ was proposed by Deardorff (1970a, b). Experiments in CBLs in Minnesota (Izumi and Caughey, 1976; Kaimal et al., 1976), England (Caughey and Palmer, 1979), France and Spain (Druilhet et al., 1983), and other places, including over-sea measurements, have confirmed the utility of this scaling for most turbulence-related quantities, including diffusion rates (for a review of the development of convective scaling and its application to diffusion, see Briggs, 1985a). The associated temperature scale, $\theta^* = \overline{w'T'}/w^*$, has also proven useful for describing the temperature profile and temperature variances in the lower half of CBLs. For a comprehensive review of knowledge of CBLs as of 1981, see Caughey (1981). Unreferenced information in the discussion below can be found in Caughey's review.

An ABL with turbulence driven mainly by buoyancy, rather than by wind shear, is considered convective. Usually, the buoyancy arises from the surface due to heating by solar radiation absorbed at the surface. However, occasionally an "inverted" CBL develops due to cooling at the top of a cloud or a fog layer produced by outgoing long-wave radiation; because the magnitude of this heat flux is considerably smaller than the typical heat flux produced at the ground by solar heating, wind speed must be fairly low for convective effects to dominate. Cloud-topped CBLs and convective mixing in fogs have received very little attention.

For the more usual CBL, with turbulence driven by heating of the surface, highly efficient vertical mixing is observed to develop; major updrafts and downdrafts extend through nearly the whole layer (Willis and Deardorff, 1976b). The following are some consequences of this vigorous vertical mixing: (1) horizontal momentum becomes well-mixed, so that there normally is negligible wind shear from $z/z_i = 0.1$ to 0.9, (2) potential temperature becomes well-mixed, but there is an increasingly negative $\partial\Theta/\partial z$ towards the surface, and (3) passive material becomes almost uniformly mixed in the vertical in a time $\simeq 4\, z_i/w^*$ after release, typically about a half hour. The CBL is capped by stable air; waves in this air and undulations in z_i, are produced by the impingement of stronger updrafts against the stable capping layer. These strong updrafts also entrain some of the warmer, stable air aloft and bring it back down into the CBL, creating a negative heat flux at the top of the CBL; this usually ranges from 0.1 to 0.4 times the surface heat flux in magnitude. Another consequence of this entrainment is that significant horizontal momentum is brought into the CBL; if there is a large difference between the wind in the CBL and the wind just above z_i, this entrainment can cause considerable wind shear in the upper part of the CBL.

In CBLs, velocity variances are most closely related to $w*$, at least when $z_i > 20|L|$. (The ratio $z_i/|L|$ is considered the index characterizing the degree of convectiveness. A $z_i/|L| > 50$ can be considered "strongly" convective; however, using a numerical model, Deardorff (1972) found significant changes from NBL turbulence structure at $z_i/|L| = 1.5$, which is considered only weakly convective.) Panofsky et al. (1977) fit measured values of both σ_u and σ_v to $(12u*^3 + 0.175w*^3)^{1/3}$. Note that $u*$ and $w*$ make equal contributions to the horizontal variances when $w*^3/u*^3 = 69$, which is equivalent to $z_i/|L| = 28$ (with $k = 0.4$). More recently, Hicks (1985) analyzed data from experiments over land and tropical seas; he recommended $\sigma_v = (3.6u*^2 + 0.35w*^2)^{1/2}$, which yields equal contributions from $u*$ and $w*$ at $z_i/|L| = 13$. At larger values of $z_i/|L|$, a very practical approximation to all measurements is $\sigma_u = \sigma_v \simeq 0.6w*$ (Briggs, 1985a). This approximation holds quite well throughout the CBL, even down to 4 m above the surface in the Minnesota measurements. This constancy with height has been convenient for the purpose of developing simple parameterizations for lateral diffusion (Briggs, 1985b); it also gives a simple way to estimate $w*$ from near-surface, fast-response measurements of u. The vertical velocity variance can have three asymptotic regimes in the CBL: $\sigma_w \propto u*$ at $z << |L|$, $\sigma_w \propto (H*z)^{1/3}$ at $z > |L|$ but $< 0.1\ z_i$ (provided that $z_i/|L| >> 10$), and $\sigma_w = w*$ times a function of z/z_i at $z > 0.1\ z_i$. For the lowest 10% of the CBL, Panofsky et al. (1977) suggest $\sigma_w = (2.2u*^3 + 2.3\ H*z)^{1/3}$. Empirical fits to field observations for the upper CBL σ_w have been suggested by Irwin (1979) and Hicks (1981).

One aspect of vertical velocities in the CBL that has considerable consequences for diffusion is that they are strongly non-Gaussian in distribution, at least through the bulk of the mid-CBL. Lamb (1981) has shown this from his numerical modeling, and field measurements also show highly skewed w distributions. For z/z_i = 1/4 and 1/2, the mode of Lamb's distributions is at $w \approx -0.5w^*$. This agrees well with the observed descent rate of average plume centerlines from elevated sources, which will be discussed later in this report. Lamb's interpretation of this, which has wide support in field observations, is that downdrafts occupy a larger percent of the horizontal area than do updrafts (thermals); the updraft/downdraft area ratio is about 0.4/0.6. Updrafts tend to have about 1.5 times as large absolute velocity, so the condition $\overline{w} \approx 0$ is satisfied. The significant skewness in w' distributions makes conventional Gaussian plume modeling unrealistic for CBLs.

Hanna (1981) has given relationships for Lagrangian time scales in the CBL; basically they are z_i/w^* times functions of z/z_i. Because updraft and downdraft velocities average about $0.6w^*$ and $0.4w^*$, respectively, the time it takes for a passive substance to traverse z_i once up and once down is approximately $4z_i/w^*$, which is about how long it takes concentrations to become almost uniformly mixed. For turbulence measurements from fixed sensors (Eulerian measurements), the peak energy content is observed at a frequency $\approx 1.3z_i/\overline{u}$ for the horizontal components and approaches $1.5\ z_i/\overline{u}$ for w' in the upper part of the CBL (Kaimal et al., 1976). This is in rough agreement with Deardorff and Willis' (1976b) laboratory observations that typical downdraft "cell" sizes were $\approx 1.5\ z_i$. One implication of these observations is that for good ensemble averaging of any quantity in the CBL, including concentration, one should average over a period many times 1.5 $z_i/\overline{u}$, to average out the effects of individual eddies; 1.5 $z_i/\overline{u}$ typically ranges from 1 to 15 min.

The height z_i begins at about the height of the nocturnal boundary layer at sunrise and increases rapidly through the morning once heat flux, or H*, becomes significantly positive (this may be delayed until most dewfall or surface wetness from precipitation has been evaporated). As a first approximation, the growth of z_i can be predicted by using an early morning temperature sounding and the time-integrated surface heat flux and by assuming that all the heat flux is accounted for in warming of the air beneath z_i from its initial temperature to a uniform potential temperature, $\theta_s(z_i)$ (subscript "s" refers to the sunrise potential temperature profile). Such a model is called an "encroachment" model. In differential form, it can be written simply as

$$z_i(\partial\theta_s/\partial z)_i \, dz_i = (\overline{w'T'})dt, \qquad (2.4)$$

where $\overline{w'T'}$ is proportional to the sensible heat flux at the surface and $(\partial\theta_s/\partial z)_i$ is the morning profile value at $z = z_i$. According to Tennekes (1984), the encroachment method handles about 80% of the z_i prediction problem; adding an additional 20% heat flux as a crude approximation of the negative heat flux at z_i adds another 10% accuracy, and further refinements in the prediction method "tend to get lost in the unavoidable inaccuracies of most experiments."

Cumulus clouds often develop just above the mean mixing depth, especially in the afternoon on humid days. These are caused by the transport by thermals of humid air from near the surface to its lifting condensation level (LCL). Vertical motion in CBLs is nearly adiabatic, so air cools nearly 1°C for every 100 m of rise. At first, only the most vigorous thermals containing the largest absolute humidities reach the LCL to form small scattered cumuli. As z_i grows, more and more thermals reach the LCL and the percentage of sky cover by cumuli increases; this phenomenum has been discussed and modeled by Wilde et al., 1985. One consequence of this mechanism in very humid weather is that, at some point during the day, nearly the whole sky becomes covered with cumulus clouds, which limits the solar radiation reaching the ground and, in turn, limits further growth of z_i.

Depending on conditions above z_i, "cloud venting" (Ching et al., 1983) may occur, causing transport of air and pollutants out of the mixed layer into the overlying stable air. This venting occurs when "forced" clouds (Stull, 1985) are carried by their vertical momentum past z_i to a level of free convection (LFC), at which point the latent heat released by condensation of water vapor is sufficient to cause the clouds to develop positive buoyancy relative to the surrounding dry air. (The existence of a LFC depends on the surface humidity and the mean temperature profile.) When this happens, the clouds become "active" (Stull, 1985) and develop vertically into cumulonimlus clouds. The rate at which CBL air can be vented aloft by these clouds has not yet been quantified, but Issac et al. (1984) have estimated that up to 90% per hour of air below cumulus cloud bases enters the clouds in some conditions; much of this air subsides and returns to the CBL, but some significant chemical changes can occur during the wet phase.

The CBL tends to "collapse" in the late afternoon as its driving force, the positive surface heat flux, is removed. Little is known about how this occurs. (This phenomenon was discussed in the subsection on "The Diurnal Cycle and Transitional Periods.")

STABLE BOUNDARY LAYERS

Research on SBLs has lagged behind the advances made for CBLs, but much progress has been made in the last seven years. Part of the reason for delayed progress was the difficulty of developing instrumentation that responded adequately to the very low turbulence levels that occur in SBLs. In addition, the structure of SBLs is inherently more variable than that of CBLs and cannot be so easily generalized; it is not known whether the relationships that have been developed on the basis of "ideal" terrain experiments will transfer to more typical sites. Terrain slope is critical because even extremely slight slopes, on the order of 0.1°, lead to downslope gravitational forces that are as important as the large scale pressure gradient in driving the flow (Caughey et al., 1979). The angle between downslope direction and pressure gradient force also can have a critical effect. Besides these effects, the temperature and stability structure of SBLs can be considerably affected by radiational cooling, and the wind velocity profile evolves slowly through the night, rarely achieving a steady state. These effects cause the Ri profile to evolve also, so that "turbulent bursts" may occur. Short of that, large amplitude wave motion may develop at the top of a CBL, causing its height to fluctuate (Lu et al., 1983).

There has been much improvement in knowledge of SBL turbulence structure for flat sites in undisturbed conditions (no discernable wave motion). High quality experiments in Minnesota (Caughey et al., 1979) and in the Netherlands at Cabauw (Driedonks, van Dop, and Kohsiek, 1978) have made this possible. Nieuwstadt (1984a,b) has shown that many relationships that are valid for the surface layer, the lowest 10% or so of the SBL, are appxoximately valid for the whole SBL if local scaling is used. Instead of scaling with $u*$, the surface value of $\tau^{1/2}$, we scale with the value of $\tau^{1/2}$ at the height in question. Likewise, we use the local value of heat flux and a local value of the Obukhov length, $\Lambda = -\tau^{3/2}[k(g/T)\overline{w'T'}]$. The reason that local scaling works is probably because eddy sizes are limited in stable conditions to a small fraction of the total SBL thickness (this is not true for CBLs or NBLs). Thus, each layer responds only to the immediately adjacent layers and not to surface fluxes. One of the useful results of Nieuwstadt's data analysis is that $\sigma_w \simeq 1.4\tau^{1/2}$ throughout the CBL. For the eddy viscosity of momentum, K_m, he found that $K_m = \tau^{1/2}\Lambda f_1(z/\Lambda)$ for the whole layer, in contrast to the expression usually assumed for just the surface layer, $K_m = u*Lf_2(z/L)$ (K_m is defined as $\tau \div \partial\bar{u}/\partial z$ and f_1 and f_2 are functions).

Nieuwstadt (1984a) also developed a simple prediction for the profiles of the local fluxes by using the equations for $\partial\theta/\partial t$, $\partial\bar{u}/\partial t$, and $\partial\bar{v}/\partial t$ in horizontally homogeneous boundary layers. To do this, h assumed (1) stationarity, with time derivatives of $\partial\theta/\partial z$, u, and v set equal to zero, (2) a constant flux Richardson number, defined by $R_f = (g/T)\overline{w'T'}/(\tau\ \partial\bar{u}/\partial z)$ (vector values, rather than scalar values, are used for τ and $\partial\bar{u}/\partial z$), and

(3) $h = 3^{1/4}(kRi_f u^*L/f)^{1/2}$. Assumption 1 is not really valid for the mean quantities like $\overline{u}$ and θ, but it seems to give good predictions for turbulence quantities (i.e., turbulence may adjust quickly to changes in gradients of mean quantities). The value of R_f, like Ri, increases from zero at $z = 0$ to a nearly constant value $\simeq 0.2$ through most of the SBL. Assumption 3 agrees with earlier arguments for the functional form of h, to be discussed later; however, a simple solution for τ is obtained only with this particular choice of constants, for which $h = 0.35\ (u^*L/f)^{1/2}$. The solution obtained is $\tau/u^{*2} = (1 - z/h)^{3/2}$. This was shown to be in good agreement with the Minnesota and Cabauw observations. The result for heat flux, $\overline{w'T'}/(\overline{w'T'})_0 = (1 - z/h)$, follows more readily from just the stationarity assumption $\partial(\partial\theta/\partial z)/\partial t = 0$. The Cabauw observations fit this well enough, but it is seen in Nieuwstadt's (1984a) Figure 10 that the Minnesota observation would better fit a higher power in $(1 - z/h)^n$, with $n \simeq 2$ instead of 1 (this difference in sites could be due to differences in radiation flux divergence). Some consequences of Nieuwstadt's forms for τ and $\overline{w'T'}$ are $\Lambda/L = (1 - z/h)^{5/4}$ and $\sigma_w/u^* = 1.4(1 - z/h)^{3/4}$.

Lateral variances for Cabauw were not reported, but it might be noted that the σ_u/u^* values for Minnesota reported by Caughey et al., (1979) give a fairly good fit to $2.3(1 - z/h)^{3/4}$. Hanna (1981) suggested a linear decrease with height in σ_w, σ_u, and σ_v as empirical fits to the Minnesota data, with, for instance, $\sigma_v/u^* = 1.3(1 - z/h)$. However, expressions like these with σ_u and σ_v of the same order as σ_w can only represent SBLs in their most ideal state, with only small-scale, nearly isotropic turbulence and no large horizontal eddies or meanders. Thus, they should be interpreted as minimum values; on some nights, at some sites $\sigma_v >> \sigma_w$ occurs.

Length scales for the peak turbulence energies observed in Minnesota were reported by Caughey et al., (1979). Hanna (1981) fit these with power laws in z and h and used these to derive semi-empirical expressions for the Lagrangian time scales. Venkatram et al. (1984) fit diffusion observations of σ_z in SBLs with statistical theory using a Lagrangian time scale $T_L = \ell/\sigma_w$ and a length scale given by $\ell^{-1} = 2.8/z + 3.7N/\sigma_w$, where $N = [(g/\theta)\partial\theta/\partial z]^{1/2}$. The first term on the right represents a neutral asymptote for ℓ, and the second term gives the effect of stable stratification; N is known as the Brunt-Vaisala frequency, a frequency characteristic of gravity waves in stable air; the maximum vertical excursion of turbulent eddies is $\propto \sigma_w/N$ in stable stratification.

Because the magnitude of turbulence depends on z/h, and there is normally no turbulence above z = h, the height of the SBL has been a subject of great interest in recent literature. Before 1973, h had not been suitably measured. It was sometimes confused with the top of the inversion layer, h_i, defined by $\partial T/\partial z > 0$ and $\partial\theta/\partial z > 0.01°C/m$ at $z < h_i$; however, in the upper part of a SBL and above, $\partial\theta/\partial z$ is strongly influenced by radiational cooling, which depends on the water vapor and CO_2 profiles among other things. Also, turbulent exchange tends to reduce $\partial\theta/\partial z$, not increase it, so $\partial\theta/\partial z$ greater than some threshold value makes no sense as an indicator of the top of the turbulent boundary layer, h. Other doubtful measures of h are the height of significant cooling, h_θ, and the lowest height of a maximum in the wind speed profile, h_u (it is true that a "jet", a wind speed maximum, often develops above the SBL, in response to inertial forces free of frictional drag). However, there is one mean profile quantity that is a good indicator of the height of turbulent mixing, and that is the Richardson number, Ri (Eq. 2.2), or the Bulk Richardson number, $R_b = (g/\theta)(\theta - \theta_0)z/\overline{u}^2$, where θ and $\overline{u}$ are measured at the height z (R_b does not require the precision measurements that Ri does, so it is often the more practical choice). By using data from 25 nights of the Wangara meteorological experiment in Australia, Mahrt et al. (1982) showed that the height at which Ri = 0.5 correlated highly with u* and, by implication, with L, as u* and L were also highly correlated (Mahrt and Heald, 1979). In contrast, the correlations between h_u or h_θ and u* were about half as large, about 0.42, and the correlation between h_i and u* was even less. They noted that Ri was normally small and relatively constant with height in a layer near the ground, but that this layer was capped by a sharp increase in Ri. In the same journal issue, another paper using the Wangara nocturnal data appeared by Wetzel (1982). He showed that the height at which $R_b = 0.33$ correlated well with u* and L, with poorer in correlations for h_θ, h_u, and h_i in place of R_b. He also found another profile height index that correlated with u* and L almost as well as the R_b index, namely the top of the linear part of the θ profile (these profiles were unusually regular, however, which may be more characteristic of ideal, flat sites like Wangara). Arya (1981) showed that u* and L correlated fairly well (r = 0.63) with h estimated from acoustic sounder returns at Cabauw, whereas the correlation of this h with h_θ was only 0.28 and with h_u was not even statistically significant.

Ideally, we would like to be able to estimate h on the basis of surface parameters, so that something simpler than a well instrumented tall meteorological tower would do the job. Zilitinkevich (1972) developed a simple SBL model with

$$h = b(u^*L/f)^{1/2} \tag{2.5}$$

This equation has received much testing in recent years. While this h correlates significantly with measured h, it is far short of perfect. For acoustic sounder measurements at Cabauw, Arya (1981) found a correlation of $r = 0.67$, while Nieuwstadt (1984a), who excluded weakly stable cases, found $r = 0.77$. Nieuwstadt and Tennekes (1981) found better correlation with a prognostic model; h never reached its equilibrium value because the relaxation time was too long (this time depended on the surface cooling rate, $\partial\theta_0/\partial t$, and became ~ 10 hr late at night). However, according to Nieuwstadt (1984a), part of the success of this model was due to the fact that the initial h was specified; because time scales are long, predicted and observed h only slowly diverge. Because there is no ready way to specify the initial h, he concludes that diagnostic models like Eq. 2.5 are more practical for now. However, Eq. 25 is a steady-state solution, and the SBL is so slow to adjust to a balance of forces that a steady state may not often be achieved. A second-order closure numerical model developed by Brost and Wyngaard (1978) gave predictions that fit Eq. 2.5 well, but the steady state, with $b \approx 0.4$, was not approached until $t = 5$ hr after the onset of surface cooling. At $t = 2$ hr, $b = 0.7$ was the best fit. The measured values of h from the Minnesota experiment runs, which were centered at about 1 hr after transition to negative H*, fit Eq. 2.5 with b = 0.7 to 0.9 when h was defined as the height at which heat flux extrapolated to only 5% of its surface value (Caughey et al., 1979).

Garratt (1982) compared Eq. 2.5 with h data from four different sites, Koorin and Wangara in Australia (both using the Ri method), Cabauw (acoustic sounder), and Minnesota (heat flux profile). By using a different criterion for h than did Caughey et al., (1979), he calculated a best fit of b = 0.37 for Minnesota; for Cabauw and Wangara, he calculated b = 0.42, and 0.39, respectively, which supports the generally accepted value of b = 0.4. However, for the only low-latitude site, Koorin, at 16°S, the best fit was b = 0.13. This is three times smaller than the other sites; the $f^{-1/2}$ term in Eq. 2.5, even if incorrect, does not account for this difference (remember that f contains the sine of latitude). Garratt accounted for this difference by running Brost and Wyngaard's model for each site while including the significant terrain slopes at Koorin (0.11°) and Minnesota (0.08°) and the angular difference between the fall line and mean surface wind direction; the results of these model runs gave very good agreement with the best-fit values of b observed at each site. This result should make us cautious about the general use of a surface parameter approach like Eq. 2.5 until it is validated for sites with more typical slopes; certainly, these often considerably exceed 0.1°.

There are diagnostic alternatives to Eq. 2.5. For sites where there are significant slope flows, gravitational acceleration may be much stronger than Coriolis acceleration. (This only requires that $g(\Delta\theta/\theta)\beta \gg \overline{u}f$, where $\Delta\theta$ is the potential temperature difference between the drainage flow and "free" air and β is the terrain slope, in radians. If $\Delta\theta = 5°C$, $\theta = 300°C$, $u = 3$ m/s, and $f^{-1} = 10^4$ s, $\beta >> 0.1°$ implies that gravitational force dominates.) In such cases, $u*/f$ should not be a significant length parameter. Instead, it may be found that $h \propto L$, as Kantha and Long (1980) found to be true in mechanically-stirred stratified fluid in a laboratory tank. For the simple two-dimensional slope flow model that Briggs (1981) fit to Ellison and Turner's (1959) laboratory slope experiments, $h \propto x$, the distance down the slope, and the mean flow speed $U \propto (H*x)^{1/3}$, this gives $h \propto U^3/H*$ (the dependency of h and U on β cancel out in this expression). If $U \propto u*$, then $h \propto L$ for slope flow; we might expect something like surface-layer similarity profile assumptions to apply to a slope flow, in which case constant h/L would imply a nearly constant U/u* (influenced slightly by a small increase in $\ln(h/z_0)$ down the slope), so that there is an internally consistent argument for $h \propto L$. Idealized as this model may be, it is adequate warning that L may be highly variable across rolling or hilly terrain at night.

In sorting through these alternatives, one should be aware of strong correlations between u*, L, and other SBL parameters; these correlations also provide possibilities for useful simplifications. Van Ulden and Holtslag (1983) showed that the surface-layer temperature scale, $T* = \overline{w'T'}/u*$, is nearly constant over a wide range of u* at Cabauw (see Fig. 2b), becoming much smaller only at very small u*, < 0.1 m/s; this constant T* does depend somewhat on cloudiness. Some consequences of constant T* are $H* \propto u*$, $L \propto u*^2$, and, if Eq. 2.5 holds, $h \propto u*^{3/2}$. Venkatram (1980) showed that

$L = 1100\ u^{*2}$ (mks units) gives a good fit to field experiment data at three relatively smooth and flat sites; furthermore, $h = 2400\ u^{*3/2}$ (mks units) gives a good fit to the Minnesota measurements of h, which ranged from 30 m to 400 m. Nieuwstadt (1984a) carried this simplification one step further to show that $h = 28\ \overline{u}^{3/2}$ (mks units), where $\overline{u}$ is measured at 10m, worked about as well as Eq. 2.5 for Cabauw observations. However, the constant in this relationship probably depends on site parameters like z_0; furthermore, in drainage flows, the relationship may be more like $h \propto \overline{u}^3$, as already discussed. Briggs (1982a) suggested an easily measured replacement for L, $L_n = \overline{u}^3/R_n^*$, where R_n^* is defined like H* but with net long-wave radiation at the surface replacing heat flux. In doing this, he showed that $H^*/R_n^* \simeq 0.5(1 + S/L)^{-1}$ for the Project Prairie Grass measurements (Barad, 1958), where S = 15 m, a site-specific length constant. This implies two regimes. One is a near-neutral regime with large L and $H^* \propto R_n^*$, limited only by the radiation balance; this gives $L \propto u^{*3}$ and Eq. 2.5 reduces to $h \propto u^{*2}$ or $h \propto L^{2/3}$. For a very stable regime, with $L << S$, $H^* \propto L \propto u^{*3/2}$ and Eq. 2.5 reduces to $h \propto u^{*5/4} \propto L^{5/6}$. In between these asymptotes, there is a broad range that approximates the constant T* solutions, as suggested by Venkatram (1980). Note that this intermediate regime gives $h \propto L^{3/4}$, and for all cases we have discussed, including slope flows, a strong correspondence between h and L is predicted; namely, $h \propto L^n$ with n ranging from 2/3 (flat, near neutral) to 1 (slope flow). Therefore, h = function(L) might be a versatile approximation, even with L replaced by an easily measured quantity like $L_n = \overline{u}^3/R_n^*$.

Besides slope effects, in SBLs we must be aware of radiative cooling effects on boundary layer structure. Little work has been done on this but Garratt and Brost (1981) made a theoretical investigation by using the same second-order closure model that so successfully modeled the effects of slight terrain slopes (Brost and Wyngaard, 1978; Garratt, 1982). To do this, they had to assume some typical profiles for water vapor, CO_2, and temperature in the troposphere plus a value for the surface emissivity (percent of black body radiation emitted). Six different cases were run, including two for which radiation flux divergence (radiative cooling) was left out of the calculation. The results showed that radiative cooling has large effects on the profiles of heat flux ($\overline{w'T'}$) below 0.1h and on θ and Ri profiles near and above $z = h$. Radiative cooling also has significant effects on the profiles of $\overline{u}$ and θ in terms of surface-layer similarity theory largely because of the development of an above-surface minimum in $\overline{w'T'}$; this affects H* and L, which are defined at the surface. Without radiative cooling, σ_w above $z = h$ does not extinguish nearly as rapidly after transition to night. On the other hand, the inclusion of radiative cooling in the model has relatively little effect on $\overline{u}$, τ, $\partial\theta/\partial z$, and Ri below $z = h$, or on the evolution of u^*, R_b at $z = h$, or $b = h/(u^*L/f)^{1/2}$ (from Eq. 2.5). At $z = h$, R_b increased from approximately 0.1 at $t = 1$ hr to 0.25 at $t = 2.5$ hr regardless of initial conditions or inclusion of radiation; b increased slowly from about 0.32 at $t = 1$ hr to 0.4 at $t = 12$ hr, except that it was of the order of 10% larger in the first few hours when radiative cooling was omitted.

At present, there are no models that could possibly predict occurrences of large amplitude waves or large time-scale horizontal eddies (meanders) in the SBL. It is understood that the possibility of large gravity waves, which may "break" into turbulence or initiate a turbulent burst, exists when Ri is near a critical value (~1/4) through a substantial layer. However, it is difficult to predict the evolution of Ri through a night, as it depends on far upwind initial conditions and the structure of the humidity profile that develops as the CBL shuts down and surface evaporation and plant transpiration continue around the time of transition. DeBaas and Driedonks (1985) made a case study of a wave occurrence measured at the Cabauw meteorological tower; Ri = 0.2 to 0.3 was measured for certain wind directions on the tower at z = 80 to 160 m. Large waves were present, as could be seen in T versus time traces from the instruments at z = 120 and 160 m; also, σ_w more than doubled between 80 and 160 m, a reverse of its usual behavior. The authors developed a linearized wave model which successfully predicted the frequency range of the waves that developed, especially as seen in the turbulence spectra for w'. However, the wave magnitude could not be predicted with this model. Another episode of large wave motions near the top of a SBL was described in detail by Lu et al. (1983); this episode was apparently caused by the passage of a shallow front.

Large time-period wind direction shifts and meanders occur in SBLs; these can cause 1-hr average lateral diffusion to be many times as large as would be predicted using boundary-layer turbulence models, which describe only three-dimensional eddies are restricted in size by the ground and by stable stratification. Large two-dimensional horizontal eddies, with no such constraints on size, often exist and tend to be site specific. Hanna (1983) has summarized empirical information from a number of sites. At every site, the magnitude of the lateral wind direction variance, which directly relates to initial lateral dispersion, tends to increase approximately as $\overline{u}^{-1}$ in stable conditions. However, σ_v tends to be nearly constant (within a wide range) with $\overline{u}$ at all sites. Its average value ranges from 0.3 m/s at Porton (England) to 1 m/s at a complex terrain site in California. Hanna also presents some wind velocity observations for a site on the Snake River Plain in Idaho, a broad plain bounded by mountains. For this site $\sigma_v \simeq 0.5$ m/s for most 1-hr average values; occasionally values are much larger, but they are rarely less than 1/2 of this. Superimposed on these fluctuations were wind direction shifts with time periods ranging from about 1 to 4 h. Such large time-scale oscillations could be due to seiche effects peculiar to that kind of topography, with cold air trapped in a large basin.

SECTION 3

DIFFUSION EXPERIMENTS

How far have they done, for what sources, under what conditions?

Good experimental information is essential for building confidence in the abilities of modeling techniques to work under all proposed conditions of application. Some partial confidence develops from experience with model applications, provided that gross model failures are rare, non-critical, or simply not publicized. However, each user is concerned with only a limited range of source and meteorological parameters. Since there is no collective archive of experiences with model failures and successes, information of this nature available to a model user tends to be anecdotal and patchy. It sheds little light on the causes of model failures and gives little basis for confidence in the model for application to radically different source types or meteorological situations.

It is far preferable to test and compare model performances against data from carefully executed diffusion experiments. Such experiments should include trustworthy measurements of the diffusion pattern of a conservative tracer (one that does not lose or gain mass flux with distance) and every source and meteorological variable that might relate to variations in the diffusion pattern (variables such as wind direction, stability, source buoyancy, etc.) If these measurements are complete, then the data set is "universal" in that it provides a standard of comparison for any model, past, present, or future. Unfortunately, such complete experiments are rarely accomplished. Also, there is a seemingly limitless number of combinations of source, terrain and surface configurations, effluent characteristics, and meteorological conditions for which dependable models are desired. Each field or laboratory experiment covers only a very limited range of these combinations. Usually, only one source type is utilized, e.g., it may be a passive source of conservative tracer, a buoyant source with no stack, building, or terrain downwash conditions complicating the flow, or a source of uniform glass beads to study deposition with known source characteristics. There are always constraints on the range of meteorological variables, due to the location, the limited goals and resources of the experimenters, or factors like the undesirability of allowing the equipment to get wet. Furthermore, the diffusion measurements themselves have many limits. The sensitivity and number of sensors used puts limits on the range and density of measurement locations, and the great majority of field measurements have been confined to near-surface concentration values due to the expense and logistical difficulties of measurement in the vertical dimension.

For some of the same reasons that diffusion experiments are limited, e.g., manageability, this paper reviews only the state of the science for continuous, passive, point sources of (presumedly) conservative tracers with no complex terrain or large surface inhomogeneities, such as land/water boundaries, within the range of diffusion of concern. These limitations nevertheless permit inclusion of most earlier diffusion experiments and quite a few recent ones.

FIELD EXPERIMENTS

There have been a few dozen field experiments on ABL diffusion that can be considered of research grade. While this seems a substantial number, they all have many limitations, and many "data gaps" continue to exist.

A few of the earlier experiments, such as Högström's (1964), used plume photography to determine plume geometric variables, such as the vertical standard deviation, σ_z, and the lateral standard deviation, σ_y. This is a valid and relatively inexpensive method for characterizing observed plume geometry, for both "instantaneous" or "relative" plume diffusion (i.e., diffusion relative to the instantaneous plume centerline) and for "total" or "time-averaged" diffusion (Nappo, 1984). This method is limited to the range that the plume can be seen against the background; depending on the plume opacity and the background conditions, this may be tens of kilometers in very stable conditions or only a few tenths of a kilometer in convective (unstable) conditions.

The vast majority of the field experiments use a metered release of a specific gas or some type of small particles as a "tracer". This is sampled using aspirated boxes known as "samplers" or, in the case of some fluorescent particles, sticky surfaces which catch the particles by impaction. A considerable number of these receptors are set out in an array intended to define the plume. Usually, they are set out in near-surface "arcs" of constant or approximately constant distance from the source with only a few degrees of azimuth between them. In a few of these experiments, attempts were made to define the vertical concentration profile of the plume by setting up a vertical array of samplers. This is always a more difficult and expensive feat, and is often frustrating. Because the towers or tethered balloon cables supporting the vertical array of samplers are very sparse in the horizontal plane, they often fail to define the plume's vertical dimension adequately, or they even miss the plume altogether. Rarely has such an array extended through the whole mixing layer, so its usefulness is confined to near the source where vertical diffusion is still within its bounds. Very recently, diffusion experiments have been made that use remote sensors, especially lidar, that effectively solve the problem of measuring concentrations in the vertical dimension; two of these experiments will be described at the end of this section.

There have been several good surveys of diffusion field experiments up to about 1978, which this discussion will exploit. Horst et al. (1979) selected ten field experiments that they considered adequate for testing predictions of σ_y and σ_z by state of the art diffusion theories, especially for the statistical theories, which rely on good measurements of lateral and vertical wind direction fluctuations. They carefully evaluated each experiment by considering the effects of site topography and surface characteristics, the height of release, variations in sampling times, and characteristics of the wind measurement instruments. Draxler (1984) wrote a much broader survey of diffusion and transport experiments about the same time (its publication was considerably delayed). Emphasis was on controlled-source "pure" experiments that avoided effects due to buoyancy, downwash, deposition, etc.; however the survey's scope was somewhat wider than that of the present report, and it included diffusion experiments in complex terrain, near shorelines, and within forests. Draxler's survey includes 20 "classical" diffusion experiments, made over homogeneous terrain, and 32 experiments in other categories. Another survey of some relevance was made by Sklarew and Joncich (1979); it covered field programs plus monitoring networks at power plants. Although these sources were all buoyant, buoyancy effects dominate lateral diffusion only at small to moderate distances. At tens of kilometers, plumes from buoyant sources diffuse much like passive plumes released from the same effective source height or from any height within a well-developed convective mixing layer, which tends to mix all material nearly uniformly throughout its depth.

Classic Tracer Experiments

It is not possible in a few pages to discuss individually each field experiment falling within the scope of this paper. Instead, information from Horst, et al. (1979), Draxler (1984), and also from tables published in Draxler (1976) and Irwin (1983) have been used to construct the summary of experiments shown here in Table 1. Included are four experiments that Draxler (1984) considered non-classical because of rough (but not "complex") terrain, urban siting, or low wind speed conditions; these are the Julich and Karlshruhe experiments, the St. Louis experiment, and the INEL, Idaho experiment, respectively. For these experiments, the sites seem reasonably homogeneous, at least among typical "real world" sites, as opposed to ideally flat and uniform sites like that of the Prairie

Table 1. Summary of Tracer

Site or Program	Terrain/Roughness Length (cm)	Stab.[a]	Release/ Receptor Heights (m)	Dur.[b] (min)
Surface Source				
Round Hill	Low hills, <30m/>10	U-S	0.3/2	10
Prairie Grass	Flat prairie/0.6	U-S	0.5/1.5[+]	10
NRTS, Idaho	Flat, sagebrush/1.5	U-S	~1/~1.5[+]	60
Ocean Breeze	Low dunes, <6m	U-S	2-3/1.5-4.5	30
Dry Gulch	Sloping mesa, ravines	U-S	2-3/1.5	30
Green Glow	Flat, sagebrush/3	U-S	2.5/1.5[+]	30
Hanford-30	Flat, sagebrush/3	U-S	2.5/1.5[+]	60
St. Louis	Urban, <30m relief	U-S	sfc, bldg[m]/sfc	60
INEL, Idaho	Flat, sagebrush/1.5	vS	1.5/0.8[+]	60
Hanford-67	Flat, sagebrush/3	N-S	2/1.5[+]	10-30
Elevated Source				
Brookhaven	Fields, woods, hills/100	U-S	108/sfc	60
Porton, ENG	Rolling downs/5	sU-sS	140-255[q]/cable	30
Hanford	Flat, sagebrush/3	S	56/1.5[+]	15-60
NRTS, Idaho	Flat, sagebrush/1.5	U	46/1[+]	30
Agesta, SWED	Wooded hills <50m/60	N-vS	50/photos	60
Suffield, CAN	Rolling prairie/2	U-S	7 & 15/0	30-60
Julich, GER	Forest, farm/40 to 80	U-S	50 & 100/sfc	60
Karlsruhe, GER	Forest, farm/110	N	100/sfc 20-30	
Hanford-67	Flat, sagebrush, 3	N-S	26 & 56/1.5[+]	10-30
Cabauw, NETH	Flat meadows, trees/10-20	sU-N	80 & 200/sfc	60
Copenhagen, DEN	Urban residential/60	U-N	115/sfc	60

Footnotes: [+] Plus vertical profiles. [a] Stability range, U = unstable, N = neutral, S = stable, s = slightly, v = very. [b] Duration of release or of sampling. [c] All include $\bar{u}$ or $\bar{u}$ profile. [d] Deposition may significantly reduce concentrations at the larger arc distances (Gryning et al. 1983). [e] Six 17.5-m towers at $x = 100$ m. [f] Wind and temperature, surface to specified height. [g] Profiles of $\bar{u}$ and T to 16 m, aircraft soundings to 3 km, solar radiation, soil temperature profiles and more. [h] Significant deposition noted. [i] Five 30-m towers at $x = 400$ m. [j] Significant deposition loss (Simpson, 1961). [k] Reliability questioned by Horst et al. (1979).

References for Table 1: Round Hill, Cramer, Record and Vaughan (1958). Prairie Grass, Barad (1958) and Haugen (1959). NRTS, Islitzer and Dumbauld (1963). Ocean Breeze and Dry Gulch, Haugen and Fuquay (1963). Green Glow and Hanford-30, Fuquay, Simpson, and Hinds (1964). St. Louis, Pooler (1966) and McElroy (1969). INEL, Sagendorf and Dickson (1974). Hanford-67, Nickola (1977). Brookhaven, Smith (1956). Porton, Hay and Pasquill (1957). Hanford, Hilst and Simpson (1958) and Elderkin

Experiments in Homogeneous Terrain

Tracer	Diffusion Data	Maximum Range (km)	Meteorological Data[c]
SO_2[d]	3 arcs	0.2	σ_a, σ_e, 2m; ΔT, 12-1.5m
SO_2[d]	5 arcs, towers[e]	0.8	σ_a, σ_e, 2m; profiles[f] 16m, etc.[g]
Uranine dye[h]	6 arcs, towers[i]	3.2	σ_a, σ_e, 4m; ΔT, Ri, 16-4m, 8-1m
ZnS particles[j]	3 arcs	4.8	σ_a[k], 3.7m; ΔT, 16-2m
ZnS particles[j]	5 arcs	5.7	σ_a[k], 3.7m; ΔT, 16-2m
ZnS particles[j]	6 arcs, towers[l]	26	σ_a, 2m; Ri, 12-1m; profiles[f] 200 m
ZnS particles[j]	5 arcs, towers[l]	13	σ_a, 2m; Ri, 15-2m; profiles[f] 200m
Fluores. part.	3 or 4 arcs	8, 16	σ_a, 39 & 140m; ΔT, 140-39m; clouds
SF_6	3 arcs, towers[n]	0.4	σ_a, 4m; ΔT, 32-8m; profiles[f] 61m
Four kinds[o]	9 arcs, towers[l]	13	σ_a, 1.5m; ΔT, 15-1m; profiles[f] 200m
Oil fog	Width, X/Q	6	gustiness[p]; ΔT, 124-15m
Lycopod. spores	σ_z-cable[r]	0.5	σ_e, release ht; ΔT, 7-1m; clouds
Fluores. dye	σ_y[d], σ_z-cable[r]	1.5	σ_a, 61m; Ri, 76-15m; profiles[f] 120m
Uranine dye[h]	6 arcs, towers[i]	3.2	σ_a, σ_e, 43m; ΔT, 43-3m
Oil fog	σ_y[s], σ_z[s]	5	ΔT, 122-30m; profiles[f] 122m
Glass beads[t]	σ_y-deposition	1.1	σ_a, σ_e, 8 & 16m; ΔT, 4-0.5m
Cu^{64}, Ho^{166}[u]	sparse arcs	10	σ_a, σ_e, 50 & 120m; ΔT, 120-20m; insol
HTO, HC[v]	sparse arcs	4	σ_a, σ_e, 100m; ΔT, 100-30m
Four kinds[o]	9 arcs, towers[l]	4	σ_a, 26 & 56 m; ΔT, 15-1m
SF_6	1 arc	~2	σ_a, σ_e, profiles$_f$ 200m; fluxes[w]
SF_6	3 arcs	~6	σ_a, σ_e, 115m; profiles[f] 115m; z_i

[l] Five towers each arc, 27 m high at x = 200 m, 42 m at 800 m and 62 m at 1600 and 3200 m. [m] Some releases from top of 3-floor building. [n] Eight 9-m towers at x = 200 m. [o] ZnS, uranine, rhodamine B, Krypton-85; deposition believed to be significant beyond 0.2 km (Horst et al., 1979). [p] Correlates with σ_a. [q] Line source release. [r] Samplers on cable tethering balloon. [s] Both relative to axis and total, photographic method. [t] Spheres with 14-58 cm/s settling velocity. [u] Sulphate aerosols. [v] Tritiated water vapor, halogenated hydrocarbons. [w] Plus rawinsondes, acoustic sounder, radiation, geostropic wind, z_i.

---

et al. (1963). NRTS, Bowne (1960) and Islitzer (1961). Ågesta, Högström (1964). Suffield, Walker (1965). Jülich, Vogt and Geis (1974). Karlsruhe, Thomas et al. (1974) and Thomas and Nester (1974). Hanford-67, Nickola (1977). Cabauw, Nieuwstadt and van Duuren (1979) and Agterberg et al. (1983). Copenhagen, Gryning (1981) and Gryning and Lyck (1984).

Grass experiment (Barad, 1958). A scan of column 2 of Table 1 shows experiments made in a wide variety of site types. For completeness, the more recent experiments made in Cabauw (Nieuwstadt and Van Duuren, 1979) and in Copenhagen (Gryning and Lyck, 1984) are included; both experiments used elevated releases of SF_6, which has been favored in recent years as a very conservative tracer that can be accurately measured.

Table 1 cannot be considered an adequate, or even absolutely accurate, description of each field experiment; it is very compact and omits many details. Yet it does serve to give an overall view of what has and has not been studied in the field in scientific diffusion experiments. Within the surface source and elevated source categories, the listings are approximately chronological. This ordering shows a trend toward more mass-conservative tracers such as SF_6 in more recent experiments.

There is evidence of deposition losses for almost all the tracers used in experiments before the early 1970s. Certainly, this can be suspected for any of the particle or dye tracers, for surface releases and for elevated releases after substantial surface impact occurs. Deposition may not affect lateral diffusion, or σ_y, very much, but it can greatly distort vertical diffusion at distances of about 1 km or more. For two of the surface source experiments, it was possible to estimate the decrease in mass flux of tracer using an arc of samplers on towers. In the Green Glow experiment, which used zinc sulfide particles, the loss of source material at $x = 3.2$ km was estimated by Simpson (1961) to range from 78 to 97%, with 18 to 45% loss even at $x = 0.2$ km. The greatest loss was for the most stable run. Similarly, 10 to 54% losses of uranine dye at

$x = 0.4$ km were estimated for the NRTS experiments (Islitzer and Dumbauld, 1963). Significant deposition was also suspected for the multiple-tracer releases of the Hanford-67 series beyond $x = 0.2$ km (Horst et al. 1979). Even SO_2 gas is known to deposit on the ground; Gryning et al. (1983) estimated a 20 to 25% reduction in surface concentration at $x = 0.2$ km for the Prairie Grass experiment. At the farthest arc, $x = 0.8$ km, they estimated a 25 to 45% reduction, the greatest reduction being in stable conditions. The significance of these deposition losses is that we must regard all estimated vertical dispersion results from the surface source experiments with suspicion or at least as in need of careful interpretation. The reinterpretation of the Prairie Grass experiment by Gryning et al., if correct, implies that all previous determinations of σ_z from this experiment at $x = 0.8$ km in stable conditions may be about a factor of 1.8 too large, since they were based only on surface samples and the assumption of a Gaussian, conservative plume. Hence, for surface releases, only the INEL experiment using SF_6 as a tracer should be free of this interpretive problem (this was a rather specialized experiment, run under extremely stable conditions, with sampling to only 0.4 km).

Another tracer that should not be considered conservative is oil fog, which was used in several of the elevated source experiments. Oil fog consists of micron-sized droplets formed by condensation of oil heated to a vaporous state and mixed with air; it is produced much like steam. It contains some volatile substances, and these slowly evaporate after the tiny droplets have formed, as they have a very large surface area-to-volume ratio. Lidar scans of oil fog plumes have indicated a loss of total reflectivity with distance. Scans from the 1983 CONDORS experiment, described at the end of this section, indicate a 50% loss in the total return signal (after correction for attenuation and background haze) from the plume in travel distances of 1.4 to 2.8 km for most cases (Eberhard, 1985: personal communication). The rate of evaporation depends on the oil type, and the oil chosen for this experiment was relatively low in volatile substances; most oils would evaporate faster. Presumedly, this decrease of droplet size, mass flux, and reflectivity with distance would not much affect plume width determinations based on a certain percentage of the centerline concentration, as was used at Brookhaven (the data base for the Brookhaven σ_y and σ_z curves). However, the concentration-to-source strength ratio, X/Q, of oil fog will become increasingly lower than that for a conservative tracer; when σ_z is inferred from surface measurements by assuming a Gaussian, conservative plume, as was done at Brookhaven, increasing overestimates of σ_z with distance would result.

The Agesta experiments of Högström (1964) also used oil fog, but plume parameters were determined by photographing puffs of oil fog from near the releasing point; each puff was photographed for about 15 min, until it became difficult to define, and then another puff was released. The meandering component of diffusion was determined from the vertical and lateral standard deviations of puff centerlines as the succession of puffs passed designated travel distances (estimated from travel time and wind speed measurements); this determination should not be affected by droplet evaporation. However, the relative diffusion estimates, based on the average puff size relative to the puff center, could become increasingly too small with distance, as evaporation could cause the apparent edges of the oil fog puff to shrink somewhat. At the furthest distances, the opposite effect could occur, as the more rapid disappearance of the apparent puff could lead one to believe that it became more rapidly diluted by a more rapid increase in σ_y and σ_z.

With the possibility of "disappearing tracers" in mind, consider the limitations of the diffusion data in experiments listed in Table 1. First of all, for about half of the experiments, tracer measurements were made only within a few meters of the surface, by using arcs of samplers. In such cases, only the lateral dispersion near the surface, such as σ_y, and the crosswind-integrated concentration, $\int \chi dy$, are measured directly. The $\int \chi dy$ measurements are universally meaningful only if little of the tracer is lost at the distance of the arc, i.e., if the integrated mass flux overhead $\simeq Q$, the rate of release. If significant deposition has occurred, this measurement is relevant only for material with the same deposition rate under the same conditions. Very often, σ_z has been inferred from $\int \chi dy$ measured on surface arcs by assuming (1) a conservative tracer and (2) a Gaussian form of vertical concentration distribution about the release height. We have reason enough to believe that assumption 1 is very risky after about 0.1 to 0.4 km of plume contact with the surface for the majority of the tracers (some have yet to be proven or disproven as approximately conservative tracers). Assumption 2 has been shown to be roughly correct in neutral and stable conditions, and in unstable conditions for near-surface sources at distances less than 0.5 $z_i\bar{u}/w^*$ (Pasquill, 1974; Willis and Deardorff, 1976a). At larger distances and for elevated sources in unstable conditions, this assumption may be grossly incorrect (Willis and Deardorff, 1976a, 1978, and 1981).

The σ_z curves most used in current practice, Pasquill's and Brookhaven's (Section 4), were derived using surface sampling data with assumptions 1 and 2, above; therefore, these curves should be considered very doubtful for the unstable cases, unless they are applied to the same type (surface or elevated) of source using the same assumptions. Pasquill's non-neutral σ_z curves are based mostly on the Prairie Grass experiment, so they apply best to near-surface sources. Brookhaven's curves are based mostly on their 108-m oil fog releases, so are more appropriate for elevated sources. In both data sets, it seems likely that the mass flux of the tracers decreased significantly at the larger distances, violating assumption 1. This would have the effect of causing overestimates of σ_z at the larger distances in all stability conditions, especially for the surface source in stable conditions.

Direct measurements of vertical diffusion were made in six of the surface-source and four of the elevated-source experiments in Table 1; samplers were mounted on either towers or tethering cables supported by balloons. How much information might these experiments give on vertical dispersion, particularly σ_z? For the surface sources, there is still the problem of deposition on the surface for all of the tracers except SF_6;

this deposition would tend to deplete the lower part of the plume more than the upper part of the plume and thus distort the profile shape compared to that of a conservative tracer plume. However, surface deposition should not reduce the maximum height obtained by the tracer at any given distance; the settling velocities, or fall speeds, of the particulate tracers were negligibly small. Therefore, the σ_z values computed from tower profiles for the surface plumes should be roughly similar to those of truly conservative tracers, although they will be affected somewhat by the profile shape distortion. To take an extreme example, if the Gaussian curve centered at $z = 0$ is distorted by a factor $\propto z$, the computed second moment around $z = 0$ (i.e., σ_z) is increased by a factor $2^{1/2}$, or by 41%. The increase is probably considerably less for actual cases affected by surface deposition, perhaps around 20% at most; thus, tower-derived σ_z values may be acceptable in spite of surface deposition. The Table 1 footnotes on tower distances then suggest that usable σ_z information exists for surface sources from $x = 0.1$ to 3.2 km. However, in every case the site is flat and smooth; z_0 ranges from 0.6 to 3 cm. There are further restrictions regarding stability. For very stable conditions, we have only the INEL experiment with towers only at $x = 0.2$ km (in the most stable runs the plumes may have reached their maximum vertical growth, with σ_z only $\simeq 4$ m, as close in as 0.1 km; the surface arcs indicated no trend in $\int \chi dy$ from $x = 0.1$ to 0.4 km, and SF_6 is considered a conservative tracer). Moderately stable and neutral cases are well covered, but the towers are too short to encompass the rapid vertical development of plumes in unstable conditions.

For elevated sources, surface deposition should have negligible effect when $\sigma_z < z_s$, the source height, because surface contact is minimal up to this point. Deposition should also be less serious than for surface sources at all distances, since concentrations near the ground are nearly always lower for an elevated release. Towers were used in two experiments, but at NRTS they were only 30 m high. The source was at 46 m, so only a small portion of the plume, not enough to determine σ_z, could be measured. The towers for the Hanford-67 experiment (Nickola, 1977) ranged from 27 to 62 m in height, so they could have substantially defined the vertical profile only for the 26 m-high release at the farther arcs (with the taller towers) in the more stable cases. Nonetheless, Horst et al. (1979) were able to determine best-fit σ_z values for a Gaussian plume on the basis of the partial profiles at $x = 0.2$ km, with good agreement among results for the different tracers used. They considered this procedure inadequate at the larger distances, due to profile distortion by deposition. Two of the elevated-source experiments used impaction-type collectors mounted on tethering cables with much less restrictive heights than tower networks, but there were no elevated, lateral plume measurements (it would be unwieldy to construct an arc of tethered balloons, and they would tend to get tangled in turbulent winds). The Porton experiments (Hay and Pasquill, 1957) defined σ_z for source heights up to 255 m, but only to $x = 0.5$ km; furthermore, the stability did not range far from neutral. The Hanford experiment measured the plume released at $z_s = 56$ m for up to $x = 1.5$ km, but only in stable conditions. Thus, we have very limited direct σ_z information for elevated sources.

Crosswind-integrated surface concentration is directly related to σ_z if the profile shape (e.g., Gaussian) and the overhead mass flux are known. Most historic σ_z estimates were made by using surface sampling arc $\int \chi dy$ measurements and assuming a Gaussian plume shape and a conservative tracer, but one or both of these assumptions are questionable in the majority of cases, as discussed earlier. However, $\int \chi dy$ is useful in its own right, mainly for computing long-term averages of surface concentrations. We can consider it to be reliably determined only for those experiments using a conservative tracer, which would only be the SF_6 experiments in Table 1. This includes only the measurements to $x = 0.4$ km for a surface release in very stable conditions at INEL and two elevated-source experiments, both relatively recent and both European. (It is curious that, in Table 1, all the surface-release experiments were made in the U.S.A., while most elevated-source experiments were made in Europe.) At Cabauw, there was only one "arc", really lines of samplers following roads. In Copenhagen, three arcs were established over a useful range of x, 2 to 6 km. Neither experiment included stable cases, nor very convective cases, rare in these northern, coastal regions. It is encouraging that both of these experiments were made in "real-world" locations with surfaces typical of many industrial and power production sites.

Because σ_y is not so affected by tracer evaporation or deposition, Table 1 indicates that there is a wide variety of useful σ_y information, particularly as derived from sampling arcs. Many of these determinations extend to 4 to 6 km downwind, for both source types. For surface sources, the Hanford experiments listed in Table 1 (including project Green Glow) extend σ_y information to 13 or 26 km, for all stabilities. However, keep in mind the fact that in stable conditions, σ_y is very site dependent; gravitational flows of cool, denser air near the surface develop even on extremely slight slopes, and these collide and interact with each other, causing horizontal meander. The investigators for the INEL experiment, which was performed during very low wind speed and very stable conditions, concluded that the best "model" for horizontal diffusion at short distances was the frequency distribution of observed wind directions (Sagendorf and Dickson, 1974). One of the unique aspects of this experiment was that the arcs extended over a full 360°; during some of the one-hour runs, the plume meandered through the full 360° of arcs. Experts have been saying for years that the only reliable indicator of σ_y averages of more than a few minutes duration in stable conditions is wind variance measurements at the site at the height of release (Hanna et al., 1977); no amount of experimental information on σ_y during these conditions is going to lead to satisfactory "universal" σ_y curves. It can only serve to establish practical upper and lower bounds.

A further important aspect of Table 1 should be considered: the meteorological data that were collected during these experiments. Very few of the experiments approach meteorological completeness so that they can be compared to any reasonable modeling approach. We now know that the mixing depth, z_i, is the most important meteorological parameter besides wind speed (the mean wind speed was measured during all experiments, usually at several levels). Only the recent experiments at Cabauw and in

Copenhagen included z_i estimates from soundings. Of the historic experiments, only the Prairie Grass data tabulation of 1956 included temperature and humidity soundings to such a depth that z_i could be determined at a later date, as was done by Nieuwstadt (1980) and Briggs and McDonald (1978). Both these references showed that the Prairie Grass unstable data compared well with predictions that use convective scaling, which requires z_i, but scatter was somewhat large owing to the rather short release time of 10 min; the bulk of the experiments had 30- to 60-min releases or sampling durations.

From the point of view of the statistical theories of diffusion, the most important measurement besides wind speed is wind direction variance. The lateral direction variance, σ_a, is needed for correlations with σ_y, and the vertical directional variance, σ_e, is needed for correlations with σ_z (the subscripts "a" and "e" stand for azimuth and elevation). Note that σ_a or "gustiness" of lateral wind direction was measured in almost every experiment in Table 1. However, most older wind vanes were poorly damped and had too much inertia, so they did not yield accurate σ_a measurements. Better instruments were available by the late 1960s. In about half of the experiments, but dominately the elevated source experiments, σ_e was measured; of these, the only ones with direct tower- or cable-measured vertical dispersion were Prairie Grass at 0.1 km, NRTS at 0.2, 0.8, 1.6, and 3.2 km (surface source) and 0.4 km (elevated source), and Porton, with measurements to 0.5 km. Thus, there is a rather limited experimental basis at present for comparing vertical diffusion with statistical modeling predictions.

Surface-layer similarity theory, favored by some modelers, especially for vertical diffusion, requires some estimate of the Obukhov length, L. This parameter is very difficult to measure directly, as it requires high quality turbulent flux measurements. This was done only in the Cabauw experiment and in the more recent CONDORS experiment to be discussed shortly. However, the Richardson number, Ri, is a function of $\overline{z}/L$ and can be used to determine L; Ri is defined by temperature difference and wind speed difference between two levels. It was measured and tabulated in four of the experiments, but could be derived for any of the experiments having high quality profiles of temperature and wind. In addition, L can be inferred from surface roughness and the bulk Richardson number by using known flux/profile relationships from surface-layer similarity theory. This number requires only ΔT between the two levels and $\overline{u}$ at one level to be determined; such measurements were made for every experiment in Table 1, so it is possible in principal to make some estimate of L. However, precisely calibrated measurements of T are required to accurately measure ΔT; often these measurements are not sufficiently reliable. Another limitation on this approach is that the surface layer may be quite shallow in moderately to very stable conditions, so for the flux/profile relationships to be valid, ΔT and $\overline{u}$ must be measured rather low, preferably below $z = L$; this condition was not met for many of the elevated source experiments.

It is also desirable to compare diffusion observations with the more empirical prediction schemes that use Pasquill/Turner categories or ΔT. While every experiment included either temperature profiles or ΔT, any comparison using ΔT would be confounded by the great variety of height pairs used to define it; experiments including detailed temperature profiles provide the possibility of standardization. Either cloudiness or solar insolation measurements are necessary to determine Pasquill or Turner categories for each run. This information was provided in only five of the experiments; this seems a serious oversight in the other experiments, considering the ease of these particular measurements and the popularity of the Pasquill/Turner method. It may be possible to assign a Pasquill category to these experiments by working backwards through other schemes that relate L, Ri, R_b to these categories (e.g., Golder, 1972). Such indirectness should not be necessary. Unfortunately, too many past diffusion experiments seem designed to test only one or two modeling approaches rather than all of them.

Long-Range Measurements

As might be expected from a simple extrapolation of trends in Table 1, as needs for diffusion data extend out to 100 km or more, the experimental information tapers off rapidly. This is especially so for vertical diffusion data, which are virtually nonexistent beyond $x = 3$ km. In convective conditions, which normally prevail during the daytime, this is not a great lack because it is well established that tracers and light particulates become nearly uniformly mixed from $z = 0$ to z_i by vigorous vertical turbulence. Above z_i concentrations fall off rapidly to near zero at $z \simeq 1.1\ z_i$. For passive tracers, this vertical concentration distribution is approximately established at a time of travel $\simeq 3\ z_i/w^*$, typically 20 to 40 min; x is then typically 2 to 10 km. In neutral and stable conditions, theoretical models suggest that complete vertical mixing through the turbulent boundary layer takes a time of the order of f^{-1}, where f is the Coriolis parameter. This is about 3 h at middle latitudes, far beyond the usual travel times associated with vertical diffusion measurements.

When one considers diffusion over days of travel, daytime mixing to the height z_i is so rapid compared to vertical mixing in stable conditions that it is usually adequate to assume that vertical diffusion is "frozen" from about sunset until the convective mixing layer develops through each layer the following morning, as long as there is no appreciable settling velocity or scavenging by precipitation. When cumulus clouds develop in daytime with bottoms near z_i, material can be drawn up into them by the updrafts supporting the cumuli, a process known as "cloud venting" (Ching et.al., 1983). This can take material out of the boundary layer; if the material has a very long residence time, over a period of many weeks it will mix through the whole troposphere, where the earth's weather takes place. On these time scales, we inevitably are faced with processes that are beyond the scope of this paper, which is limited to "simple" boundary-layer diffusion. To stay within this scope, we shall limit the travel times under consideration to about 1/2 a diurnal cycle (12 hours), or travel distances to a few hundred kilometers.

For lateral dispersion over such distances, somewhat more than a dozen sets of observations have been summarized by Pasquill and Smith

(1983) and by Draxler (1984). These observations are nowhere near as substantial and well-documented as those in Table 1 for short ranges. Some long-range observations are from a classified project (Slade, 1968) and some are unpublished. Most of the published observations have very little meteorological information other than broad designator terms like stable and neutral. Because many of these observations were made by traversing a plume at a series of distances and heights with a single aircraft, and there was inadequate time for repeated traverses over such distances, they were really quasi-instantaneous measurements of plume width or of σ_y, rather than a time average. Traditional statistical theory predicts that instantaneous plume widths approach time average widths at large travel times compared to the Lagrangian turbulent time scale, but this notion has been increasingly questioned simply because horizontal eddies in the atmosphere at any time scale up to days can be found (synoptic "highs" and "lows"). In one of the unpublished experiments (Ferber and List, 1973) referenced by Draxler (1984), both instantaneous aircraft measurements and time-averaged surface sampling of a 3-hr release of SF_6 were made for the same 76-m high release in windy, neutral conditions; the time-averaged surface-measured σ_y values were 3 1/2 times as large as the aircraft-measured σ_y values in the range of 6 to 18 km downwind. For 1-hr averages at $x \sim 100$ km, the differences are probably much less, but there appear to be no data for testing this assumption.

The σ_y versus x data for the x = 2- to 100-km range summarized by Draxler (1984) all show similar power laws of growth, with n in $\sigma_y \propto x^n$

ranging from about 0.7 to 0.9. This seems to hold regardless of stability or of whether σ_y was from instantaneous aircraft traverses or from surface sampler measurements of a 3- or 4-hr tracer release. Similarly, Pasquill and Smith (1983) fit a diverse collection of data in the range x = 3 to 1000 km with n = 0.875. These are significantly larger values of n than the value 0.5 predicted by statistical or eddy diffusivity theories at large x. Draxler (1984), Pasquill and Smith (1983), and many others have suggested that wind direction shear may be responsible for the more rapid growth of σ_y at large x. However, the large-x experiments so far seem to lack wind profiles and other supporting meteorological measurements for testing models of plume growth, so we cannot progress much beyond empirical power laws.

Useful Buoyant Plume Observations

Although this paper does not cover buoyancy effects and will pass over diffusion data from buoyant sources, such as those surveyed by Sklarew and Joncich (1979), diffusion of plumes from buoyant sources is not always dominated by buoyant effects. Thus, sometimes observations from buoyant plumes can be used to test aspects of passive diffusion models, especially when atmospheric diffusion is rapid, as in unstable conditions. For instance, Carras and Williams (1983) determined vertical plume growth versus distance by using time lapse photography at four stack sites in Australia during convective conditions. For the three sources of lesser buoyancy, good agreement was found with Lamb's (1979) convective scaling predictions for passive diffusion; the most buoyant source had enhanced growth at small x, but its plume growth also fit the passive plume asymptote at larger distances (>1 km). For σ_y, Briggs (1985b) demonstrated that buoyancy has little effect on plume growth in convective conditions if the dimensionless parameter $F^* \leq 0.06$; F^* can be defined as (w^*/U) times the ratio of the stack buoyancy flux to the buoyancy flux from a surface area πz_i^2, owing to surface heating and evaporation. Thus, F^* is a measure of the relative buoyancy contained in a segment of bent-over plume compared to that in atmospheric thermals. Except for large sources of buoyancy in low wind speed conditions, the $F^* \leq 0.06$ criterion is usually met. Briggs (1985b) also presents evidence that suggests that buoyant plumes trapped within a convective mixing layer become uniformly mixed in the vertical, just as passive plumes do, when $X = (w^*/U)x/z_i \geq 30 F^*$ (or, equivalently, $x \geq 30 F_b/w^{*3}$), where F_b is the stack buoyancy parameter. Thus at these large distances, the diffusion of buoyant plumes released into and trapped with a convective mixing layer will be essentially the same as passive source diffusion from any $z_s < z_i$. In such cases, plume width measurements at large downwind distances are useful even if the plume is initially buoyant; a good example of such measurements are those reported by Carras and Williams (1981), which extend from 15 to 1000 km downwind of an isolated stack, a considerable extension beyond the observation distances in Table 1.

Remote Sensing Experiments

Another type of experiment not included in Table 1 is remote sensing experiments, which are very recent. These overcome the limitations of surface-based sampling by measuring plume concentrations with reflected electromagnetic radiation. Lidars have actually been used since the early 1960s to define the rise and other geometric properties of buoyant plumes. They send out narrow pulses of monochromatic light from a laser and then record the time lag and intensity of the reflected return. Present operational lidars work best with plumes composed of ~1 μm-sized particles, which are easily produced by using an oil fog generator (other, more conservative tracers are available, but these are also more expensive). A lidar "scan" consists of a series of pulses made at a fixed azimuth with a small increment in elevation angle made between each pulse. Up to 10 pulses per second have been obtained; a mapping of plume concentrations at a half-dozen azimuth angles can be completed in 2 or 3 min with such instruments; repetition of such rapid scanning for 30 min or more produces an adequate ensemble of individual scans for defining an average plume. To infer concentrations of tracer, or rather of tracer reflectivity, requires careful processing of the return signal to compensate for range, background haze, and attenuation of the reflected light returning through the plume and background haze. This degree of sophistication is a recent achievement (Eberhard et al., 1985); earlier experiments used lidar mainly to define the edges of buoyant plumes. One limitation of lidar is that for eye safety reasons it should not scan within 15 m or so of the surface. Thus, it is best used for elevated sources or for surface releases at distances where they are diffused through several hundred meters or more. Lidar's great advantage is that it can detect sufficiently dense plumes at a 5-km or more range. This allows the mapping of diffusion through a mixing depth of 2 km or more. Thus, lidar measurements complement observations that can be made by using samplers on surface arcs and towers.

Lidar has been used in two recent passive plume experiments, one carried out in stable conditions and the other carried out in unstable conditions. Both use oil fog as the plume tracer for the lidar, which was provided and operated by the National Oceanic and Atmospheric Administration's Wave Propagation Laboratory (WPL). The experiment made in stable conditions at Cinder Cone Butte, Idaho, was primarily a complex terrain experiment, but σ_y and σ_z determinations were made by lidar at small distances, 0.1 to 0.9 km, upwind of the isolated hill (Strimaitis et al., 1983). Source heights ranged from 20 to 60 m, and meteorological measurements on a nearby 150 m tower included temperature at eight levels, three-dimensional wind speed and turbulence intensity at five levels, insolation, and net radiation. These measurements are sufficient for testing any presently known modeling approach and have been used to test a combination statistical/similarity equation for σ_z (Venkatram et al., 1984).

The CONDORS experiment (CONDORS = Convective Diffusion Observed with Remote Sensors) made at WPL's Boulder Atmospheric Observatory was designed to define three-dimensional plume concentrations in highly convective conditions by using both lidar-detected oil fog and doppler radar-detected "chaff". This tracer consists of bundles of aluminum-coated thread which are chopped into 1.5-cm lengths (half wavelengths) and ejected in an air jet. The radar can map the chaff plume for tens of kilometers and can detect concentrations as low as 10^{-5} threads/m^3 with almost no background or attenuation problems except for "ground clutter" presented by metallic objects such as power lines. The radar's lateral and vertical resolution is limited by its 0.8° beam width. The main drawback of this remote sensor is that the lightest chaff presently available has a settling velocity of about 0.3 m/s. Used in vigorous convection with vertical turbulence velocities of about 1 m/s, the chaff plume does mix up to z_i, however, its mode was observed to be displaced downward compared to the oil fog plume, with this displacement increasing with distance (Moninger et al., 1983). In half of the CONDORS runs the chaff and oil fog were released side-by-side at heights ranging from 167 to 280 m, using the 300-m meteorological tower. For other runs, except for one with both sources on the ground, the chaff was elevated and the oil fog was on the ground. In addition, the main part of the experiment in 1983 had one sampling arc at $x \simeq 1.2$ km, and SF_6 was released from the elevated release point; this made possible some comparison of χ/Q inferred from the remote sensors and surface tracer measurements (Eberhard et al., 1985). Because some of the chaff deposits and the oil fog slowly evaporates, a constant value of Q cannot be assumed, as for a conservative tracer; instead, the total flux at each distance, $\overline{u} \int\int \chi dydz$, was substituted for Q. (This is far superior to experiments with non-conservative tracers and only surface samplers, because there is then no direct measurement of the total flux of tracer versus distance or of the extent of vertical diffusion.) Meteorological measurements in the CONDORS experiment included temperature, three components of wind speed measured with sonic anenometers, turbulent variances, and turbulent fluxes measured at eight levels on the tower. In addition, solar radiation was measured and three or four rawinsonde releases were made during each of the 11 two-hour runs; z_i was determined from the rawinsonde temperature and humidity soundings, the top of the haze layer as seen by lidar, and the top of the chaff as seen by radar.

New Experimental Capabilities

It is essential that the most useful experiments have been those with a full complement of meteorological measurements. Of the historic experiments, Prairie Grass stands out, even though it was one of the first, made in 1956. This data set has been used much more than any other, largely because almost every needed meteorological parameter was measured or could be estimated from the available measurements. Some of the recent experiments like Cabauw, Cinder Cone Butte, and CONDORS have returned to this philosophy of measurement; they, too, should prove to be very versatile bases for model testing and will raise the standards for future diffusion experiments. All of the earlier experiments suffered to some degree from tracer deposition, unreliable turbulence measurements, and physical/economical limits on vertical sampling. Recent experiments have benefited from the use of more conservative tracers, much improved turbulence instrumentation, and remote sensors. These capabilities present many opportunities for dependable measurements over much greater ranges of distance, height, and stability conditions than was previously possible.

LABORATORY EXPERIMENTS

Fluid modeling experiments in wind tunnels and water channels have long been used as tools for simulating large-scale turbulent flows over and around obstacles. Confidence in these tools has increased to the point that they have been used extensively to investigate diffusion from releases on and near buildings and terrain features (Hosker, 1984). They are increasingly used to model diffusion in complex terrain, including stability effects, primarily because they are much cheaper to operate than large-scale experiments, particularly in difficult terrain. In the last few years, the U.S. EPA has supported several paired field/fluid modeling experiments on diffusion in complex terrain; these were directed towards improving modeling of diffusion in complex terrain and testing the reliability of fluid modeling in such applications (Schiermeier, 1984). With some limitations, the fluid modeling simulations have been quite successful.

In addition to being less expensive than field experiments, laboratory modeling offers control over the meteorological variables, so that both the flow and surface characteristics can be idealized. Therefore, it is an attractive alternative for investigation of "pure" turbulent diffusion, free of the complexities found in the field due to uneven terrain, uneven surface heating or cooling, uneven surface roughness, and variable winds that cause part or all of the plume to miss the samplers. In bypassing these complexities, fluid modeling is naturally unable to simulate some features of the real atmosphere. The most conspicuous missing element is plume meander due to large-scale horizontal flow structures many times larger than the mixing depth. Such eddies, mini-fronts, or lines of convergence may be caused by large-scale topographic variations, surface inhomogeneities, wind direction shears, or even cloud shadows in the real atmosphere. Straight-sided wind tunnels and water channels cannot simulate wind direction shear effects, and their walls prevent the development of any horizontal eddies larger than their width. Thus, laboratory lateral diffusion measurements correspond to the minimum lateral diffusion that occurs under the same stability conditions in the atmosphere during periods free of these larger scale effects. Laboratory σ_y values probably represent 10- to 20-min averages in the real atmosphere better than 60-min averages.

The great reduction of scales in laboratory modeling facilities also causes a great reduction in the range of turbulent eddy sizes compared to the atmosphere. The smallest eddy sizes are limited by fluid viscosity and the turbulent energy dissipation rate. In the atmosphere, the diffusive ability of very small eddies (~1 mm) is usually inconsequential, except for the mixing of chemically reactive species. The problem with very large scale reductions is that viscosity begins to smooth and distort even the largest eddies, which are the most effective agents of diffusion, so that none of the full-scale turbulent diffusion is well simulated. In the atmosphere, only the motion of eddies smaller than about 1 cm are effectively smoothed and dissipated by the action of molecular viscosity. The degree of scale reduction that avoids viscosity-induced distortions is most limited when modeling plume buoyancy or ambient stratification (stable or unstable atmospheres). This is because gravity then plays a role, and gravity cannot be adjusted. Gravitational acceleration has the dimensions of velocity squared divided by length; if lengths are reduced by a factor S, velocities such as wind speed must be reduced by a factor $S^{1/2}$ to maintain the ratio of inertial forces to gravitational forces (this assumes that full-scale density differences are maintained). The importance of viscosity is measured by the inverse of the Reynolds number, Re, which is a length scale times a velocity scale divided by the (kinematic) viscosity. Thus, Re must be reduced by the factor $S^{3/2}$.

To illustrate, suppose we wish to model a 1 km width full-scale terrain segment in a 2-m wide wind tunnel, so S = 500. A full-scale wind of 10 m/s blowing over roughness elements of size 1.5 m has Re $\simeq 10^6$. If we model neutral flow, we can maintain the 10 m/s flow speed and reduce the roughness size by the factor of 500, to 3 mm; this results in Re = $10^6 \div 500 = 2000$, which is large enough for good turbulence simulation. However, if we wish to model a stratified atmosphere, we must also reduce the wind speed by a factor of $(500)^{1/2}$ to 0.45 m/s, which reduces the Re to 90; this is submarginal, so that viscosity smoothes the flow around the roughness elements and they appear smoother to the wind than they should be. A non-practical solution to this problem is to put the whole wind tunnel in a giant centrifuge in order to increase the effective gravitational acceleration by substituting centripetal acceleration. More practical solutions are (1) to use less scale reduction, thus modeling less area, (2) to use increased density or temperature differences to increase buoyant accelerations, which is equivalent to increasing gravity, or (3) to experiment with exaggerated roughness elements to achieve a good simulation of full-scale wind speed and turbulence profiles. All of these strategies have their limits, and the bottom line is that there are limits to the scale reduction possible to successfully simulate atmospheric turbulence or turbulent diffusion. Several other scaling limitations of fluid modeling have been discussed by Snyder (1972 and 1981).

Laboratory diffusion experiments have been made in wind tunnels, water channels, and water tanks. The latter have been used to study diffusion by purely convective turbulence. Since tank walls prohibit crossflow, there can be no turbulence generated by boundary-layer wind shear. Water channels are relatively easy to stratify and are most often used as towing channels for buoyant sources or for elevated releases upwind of a terrain feature; that is, the model is towed along the length of the channel. This simulates the effect of a uniform crossflow velocity. This technique allows easy maintenance of strong, stable stratification, which is not easily maintained in a return flow mode (model stationary, water moving through channel and recirculating). It is used to study only the effects of crossflow and stable stratification, however, it does not permit simulation of ambient shear-generated turbulence. Such simulation requires a flow of fluid across a long fetch because the turbulent boundary layer that develops increases in depth slowly and is only a small fraction of the fetch. The boundary layer depth usually determines the scale-down factor. Only wind tunnels have been used to study diffusion in shear-generated turbulence. Large facilities with long fetches are cheaper to build and to operate using air as the operating fluid rather than water because water is so heavy; it is also easier to measure flow velocities and turbulence in air. Large wind tunnels used to simulate atmospheric boundary layers are generally about 2 m high, 2 to 4 m wide, and 30 m long.

Neutral Boundary Layers

Simulation of diffusion in NBLs is relatively easy to perform in wind tunnels because no heating or cooling is required and large flow speeds can be used. This helps keep model Reynold's numbers high enough to create well-developed turbulence that closely resembles full-scale turbulence. The main requirement for good boundary layer simulation is a long fetch over a rough surface. The boundary layer develops to a reasonable depth, e.g., 1 m, more quickly if turbulence is initiated through the whole depth of the tunnel at the upwind end by using an array of drag elements known as "vortex generators". Surface-release diffusion experiments in neutral flow were performed by Robins (1978) using such a configuration. The wind tunnel was unusually wide, 9.1 m, and 2.7 m high; this aspect ratio should allow freer development of horizontal eddies and lateral diffusion. Experiments were performed with two different floor roughnesses, one rather smooth to simulate "rural" diffusion (z_0 = 3 cm full-scale) and one rough to simulate "urban" diffusion (z_0 = 130 cm full-scale). The depth of the developed boundary layers at the release point was 0.6 m and 2 m, with scale-reduction factors of 1000 and 300, respectively, for a full-scale boundary layer thickness of 600 m. Measured turbulence intensities, when scaled with the friction velocity, u*, were within 5 to 10% of best estimates from field measurements. The mean plume measurements revealed nearly Gaussian lateral plume profiles and quasi-Gaussian profiles in the vertical, fitting $\chi \propto \exp(-az^n)$ better with $n = 1.7$ than with $n = 2$, the Gaussian value. The growth of the vertical depth of the plumes fit quite well a prediction based on surface-layer similarity theory, $\delta \ln(0.17\delta/z_0) = 0.22x$, where δ is the height at which the concentration falls to half its surface value. The lateral spread averaged approximately 20% less than a prediction based on statistical theory. Overall, Robins' experiment showed that quite good simulation of turbulence and diffusion, especially for vertical diffusion, in a neutral ABL can be achieved in a large wind tunnel. For lateral diffusion, growth due to wind directional shear cannot be simulated in a parallel-walled tunnel. In the atmosphere, wind shear effects may contribute substantially to σ_y at distances greater than a few tens of kilometers.

Convective Boundary Layers

For diffusion in convective conditions, both wind tunnel and water tank experiments have proven useful. Convective turbulence has been produced simply by strongly heating the floor of the tunnel or tank, which causes unstable stratification because both air and water become less dense when heated. Experiments of this type were first performed in a water tank by Willis and Deardorff (1976a, 1978, 1981); about a 30 cm deep mixed layer with stable capping was produced in a tank about 1.2 m square with plexiglass sides. The four walls, of course, eliminate all effects of wind and wind shear, but the tank has proven to provide a reasonably good simulation of purely convective turbulence; turbulent velocity variances, nondimensionalized by the convective velocity scale w* and plotted against z/z_i, were in good agreement with aircraft measurements for the vertical component, σ_w, and were about 30% lower than the atmospheric measurements for horizontal components (Willis and Deardorff, 1974). This may be due partly to the horizontal motion-inhibiting effect of the tank walls.

Diffusion experiments were made in this tank by instantly releasing over 1000 tiny (~1 mm), neutrally-buoyant oil droplets from a line source crossing the tank at various fixed source heights, z_s. These were photographed looking down the plume axis, and inventories of particle positions were made at a series of times, t, after release. To interpret these results for vertical and lateral diffusion in terms of distance downwind of a continuous, point source in a crossflow, it is simply assumed that $x = \overline{u}t$. This assumption appears reasonable, considering the fact that $\overline{u}$ is nearly constant with height in a very convective ABL, except close to the surface. Since the tank diffusion is, in effect, integrated along the longitudinal coordinate, x, in the photographs, this technique fails to account for longitudinal diffusion; Deardorff and Willis (1975) estimate this phenomenon to be significant for point sources only when $\overline{u} < 1.5w^*$, rather low wind speeds (a typical summer midday value of w* is 2 m/s).

The results of the Willis and Deardorff convective tank diffusion experiments were surprising and somewhat troubling to the diffusion modeling community because they revealed features of vertical diffusion that were at odds with assumptions common to all Gaussian models, so widely accepted and used. The near-surface releases, with $z_s/z_i = 1/15$, diffused in a Gaussian manner at first, up until $t \approx 0.5\ z_i/w^*$, but shortly thereafter the locus of maximum concentration lifted off the surface and lofted into the upper half of the mixed layer. This was accompanied by a rapid decrease in surface concentrations until nearly uniform vertical mixing from $z = 0$ to z_i was achieved at $t \approx 3\ z_i/w^*$, the limit of the experiment. (In convective ABLs, updrafts and downdrafts extend through nearly the whole layer with mean up and down speeds of $0.6w^*$ and $0.4w^*$, respectively (Lamb, 1982); therefore, the time needed for one complete "stir" is approximately $z_i/(0.6w^*) + z_i/(0.4w^*) \approx 4z_i/w^*$.) The elevated releases, at $z_s/z_i \approx 1/4$ and $1/2$, exhibited non-Gaussian vertical behaviors from the outset; the locus of maximum concentrations descended from the release heights until they reached the surface at $t \approx 2\ z_s/w^*$. Naturally, this produced higher surface concentrations than predicted by Gaussian models, which assume that the centerline of the plume remains at $z = z_s$. At larger t, the locus of maximum concentration lifted off the surface and lofted into the upper part of the mixing layer, just as had occurred for the surface releases.

These results were unexpected and stirred some controversy, particularly because convective tank experiments are a relatively new tool for simulating atmospheric dispersion. Some modelers questioned whether the observed diffusion patterns occur in the real atmosphere. Unfortunately, the answers could not immediately be found in existing field data due to many deficiencies already discussed, especially the lack of vertical concentration profiles to any substantive height. The same phenomena were observed in the numerical modeling results of Lamb (1979), which provided some confirmation. (Lamb's work followed the surface-release tank experiments and actually preceded and inspired the elevated-release tank experiments. For a history of developments in convective diffusion, see Briggs, 1985a.) Papers by Briggs and McDonald (1978) and Nieuwstadt (1980) confirmed the fact that surface concentration results from the Prairie Grass field experiments were consistent with the results of Willis and Deardorff (1976a) and of Lamb (1979), when convectively scaled, except that σ_y values in the tank were about 30% smaller than the field values (this seems consistent with the smaller horizontal velocity variances observed in the tank). Preliminary results from the CONDORS field experiment, which was largely motivated by the results of Lamb's and Willis and Deardorff's experiments, did show the descending plume behavior for a $z_s/z_i \simeq 1/2$ case, with a remarkably good comparison with the convectively-scaled tank results for crosswind-integrated concentrations (Moninger et al., 1983). (Convectively-scaled means that heights are scaled with z_i, time after release or $x/\overline{u}$ is scaled with z_i/w^*, and concentrations are scaled with $Q/(\overline{u}z_i^2)$.) Further good comparisons of the tank experiment results with some power plant plume data, the Cabauw field experiment, and CONDORS data were shown in Briggs (1985a). The convective tank experiment results are now being taken quite seriously, and many researchers have been developing mathematical diffusion models that fit these results (such models are surveyed in Briggs, 1985b).

More recently, Poreh and Cermak (1984) demonstrated that convective ABL simulations producing results similar to those of the tank experiments could be performed in a large boundary-layer wind tunnel. This has the advantage of including some of the crossflow and wind speed shear effects found in the real atmosphere. They ran these experiments using both a smooth floor consisting of a 12 m long heated aluminum plate and a rough floor, with twisted-link chains stretched across the plate. The simulation had some shortcomings, e.g., the fetch of heated plate upwind of the source was only about 1 $\overline{u}(z_i/w^*)$, which is not sufficient for full convective turbulence development (Briggs, 1985a), and the convectively-scaled vertical velocity variances were only about 2/3 of values observed in the atmosphere. Nonetheless, plume behaviors similar to those observed in the tank experiments were observed in the tunnel. The lateral diffusion rate was larger, even somewhat larger than that observed for the Prairie Grass field experiments (Nieuwstadt, 1980), in spite of the lateral width restriction of about 2.4 z_i in the tunnel. An improved convective diffusion experiment was recently completed in the same tunnel; stronger stable capping and increased dimensionless upwind fetch over the heated plate were achieved (Poreh, 1985: personal communication).

Stable Boundary Layers

Some basic diffusion experiments in SBLs layers have been done in the same wind tunnel facility at Colorado State University. In fact, a recent series of baseline diffusion measurements were made in this tunnel for every combination of (1) stable, neutral, or convective boundary layer, (2) smooth or rough floor and (3) surface source or elevated source (one or two different heights) (Cermak et al., 1983). The smooth floor configuration still developed turbulence (a tripping device in these experiments helped maintain turbulence throughout the test section of the tunnel). The effective roughness length of a smooth surface is $\propto \nu/u^*$, where ν is the kinematic viscosity of air (0.15 cm^2/s at room temperature); in these experiments it corresponded to approximately 1 cm at full scale (similar to the Prairie Grass field site). The rough floor configuration gave $z_0 \simeq 16$ cm, full scale, typical of mixed surface cover. To achieve the SBL, incoming air was preheated to 78°C before entering the test section, which had its floor temperature cooled to about 0°C. This large temperature difference helped to compensate for scale-down effects by increasing buoyancy forces. Unlike earlier experiments that achieved only slightly stable boundary layers, with diffusion slightly reduced from the neutral case (Chaudhry and Meroney, 1973), the recent experiments achieved substantial stability. Vertical diffusion at the larger distances was reduced by factors up to 5.5 compared to neutral conditions with the same tunnel configuration. The estimated values of the Obukhov length, 16 cm and 25 cm for the smooth and rough floors, respectively, are fractions of the boundary layer depth $\simeq 1$ m), that correspond to moderately stable conditions in the atmosphere (Pasquill category "F"). Measurements from this series of experiments were included in a comparison of wind tunnel σ_y measurements versus field data and statistical-theory predictions (Li and Meroney, 1984).

A recent diffusion experiment in a boundary-layer wind tunnel in Japan (Ogawa et al., 1985) achieved even stronger SBLs, with Obukhov lengths as small as 1/13 times the boundary layer depth. Vertical diffusion was sharply reduced from its neutral growth at the greater stabilities, results comparable to those of Cermak et al. (1983). A somewhat surprising result was that lateral diffusion was reduced from its neutral growth by about 20% in the moderately stable SBLs, while it was actually somewhat larger than its neutral growth in the strongly stable SBLs. Ogawa et al. attributed this growth to the development of horizontal meanders in the very stable flow, a phenomenon that field observations of lateral turbulence velocities tend to support (Hanna, 1983). The Ogawa et al. (1985) observations also showed significant departures from Gaussian vertical profile shapes for the neutral and very stable cases. These experiments were conducted only for a surface source and a smooth floor configuration.

Potential Further Uses of Laboratory Simulations

The results reviewed in this section support the idea that fluid modeling, in the laboratory, is a valid way to improve our understanding and modeling of diffusion. However, it is obvious that this tool has not been fully exploited. Considering the fact that it is much cheaper to run diffusion experiments in the laboratory, under controlled conditions, than in the field, it makes sense to use laboratory facilities as much as possible. These laboratory experiments should be supplemented by selected field experiments to provide confidence-building and to measure features that cannot be produced in the laboratory, like crosswind shear-induced diffusion. For neutral conditions, elevated releases in a well-developed boundary layer could be run by using several contrasting floor roughness. In a wide tunnel it would be informative to run experiments with a patchwork pattern of contrasting surface roughnesses (on a scale small compared to the tunnel width); this would be much more like the real world and might reveal effects of the surface inhomogeneities on lateral diffusion (surface drag inhomogeneities would cause vertical torques that might produce larger and more vigorous horizontal eddies). For convective conditions, a similar exploration could be made of the effects of surface heat flux inhomogeneities, which would provide forcing for the turbulent mixing much more like the real world than a uniformly heated aluminum plate (most places of interest have a complex pattern of different types of vegetation, different radiation balances, and different degrees of moistness and thus, large contrasts in surface heat flux). For stable conditions, it now appears possible to simulate fairly strong stable ABLs. There is a need for much more work on the effect of height of release relative to the boundary layer height for all stabilities.

SECTION 4

DIFFUSION MODELING AS PRACTICED

How well does it reflect the current state of knowledge?

THE GAUSSIAN PLUME - ITS UTILITY AND LIMITATIONS

The vast majority of "production line" diffusion models are Gaussian, i.e., they make the basic assumption that concentration distributions relative to the center of a puff or plume have a Gaussian shape. In a recent summary of scientific reviews of eight models (Fox et al., 1983), most of the reviewers "felt that they were reviewing a single, physically incomplete, Gaussian model with slight variations, rather than eight substantially different models". They also felt that these models do not represent the state of the science. The reason that Gaussian plumes have come to dominate the model market is a matter of precedence and convenience. Early diffusion models were naturally influenced by much earlier treatments of molecular diffusion. Molecular diffusion in a fluid is caused by countless collisions and rebounds among the molecules, which can be treated statistically as a "random walk" process. When concentration and diffusivity vary on scales that are large compared to the mean molecular path length, the diffusion process can be described by the "parabolic equation",

$$d\chi/dt = \frac{\partial}{\partial x_i} \left(K \frac{\partial \chi}{\partial x_i}\right) , \qquad (4.1)$$

where x_i ($i = 1, 2$ or 3) represents the three orthoganal spacial coordinates. In this context, K, the diffusivity, is directly related to the product of mean molecular path length and speed. In the absence of shear and when K is constant, the solution of this equation for a tracer released at a point at $t = 0$ is Gaussian; the standard deviations about the center are

$$\sigma = (2Kt)^{1/2} .$$

This approach was adapted to modeling diffusion in the atmosphere by replacing molecular diffusivity with "eddy diffusivity", which in analogy to random molecular motions is supposed to be proportional to the product of mean turbulent eddy size (or spectral peak) and turbulent velocity variance. Whenever the atmosphere is turbulent, this K is orders of magnitude larger than the molecular diffusivity, so the latter is thought to be negligible. The classical treatment for diffusion from a ground source using the parobolic equation with $\overline{u} \propto z^m$ and $K_z \propto z^n$ was first published by Sutton (1953), where K_z is the vertical component of K. The resulting distribution in the vertical is Gaussian only for the special case $m = n$. In general, $\chi \propto \exp(-\text{constant} \cdot z^p)$ with $p = 2 + m-n$. If $n = 1$, as it is observed to be in the neutral surface layer, and if wind shear is neglected ($m = 0$), an exponential distribution results. Values of p fitting observed surface-source concentration distributions have ranged at least from 1.15 to 3 (Pasquill, 1974, p. 205; Ogawa et al., 1985). However, Pasquill demonstrated that the practical effect on the ratios of various profile constants is very small for $1.25 < p < 2.5$, which suggests that there is no harm done in assuming a Gaussian distribution ($p = 2$) for these cases. Large errors result if a Gaussian distribution in the vertical is assumed when the locus of maximum concentration from a ground source becomes elevated; no value of p fits this case. This phenomenum has been observed in convective conditions, and has undoubtedly resulted in some erroneous diffusion modeling as a consequence of assuming Gaussian plume distributions.

There are many in which a Gaussian plume shape is an acceptable approximation. For a sufficient sampling duration, lateral diffusion near the source is Gaussian if the distribution of lateral turbulent velocities is Gaussian. This is normally the case, except when the sampling period encompasses a substantial wind shift. The same applies to vertical diffusion from an elevated source, almost to the point that significant concentrations reach the ground, but this is not valid in the convective mixing layer. For this case, it has been amply demonstrated that vertical turbulent velocities have a skewed distribution favoring downdrafts. The simple Gaussian assumption is wrong here and underpredicts maximum ground concentrations (Briggs, 1985a). The vertical velocity distribution can be fit with a double-Gaussian shape, however, and this has been assumed in some recent dispersion models. A near-Gaussian shape is often observed in lateral concentration distributions at distances of many kilometers. This is to be expected when the plume width far exceeds the width of the largest horizontal eddies because then horizontal diffusion becomes akin to a random walk process. At moderate to large distances in unstable conditions, after a travel distance of about $3(U/w^*)z_i$, passive material becomes almost uniformly distributed from the surface to the top of the mixed layer. Gaussian models can approximate this by assuming reflection at the surface and at $z = z_i$ (see Pasquill, 1976b).

Some further discussion of when plumes are and are not Gaussian is given by Pasquill (1978, pp. 6-9). He notes that power station plumes have been observed to be Gaussian only in their lower half. Weil (1983) speculates that the Gaussian assumption may perform satisfactorily in convective conditions for vertical distributions of strongly buoyant plumes, citing numerical simulations of Lamb (1982); however, the assumption works poorly for passive or weakly buoyant plumes. Strengths and weaknesses of Gaussian plume modeling have been discussed by Smith (1981). The conditions of applicability he lists are somewhat more restrictive than the ones stated here. Strengths that he lists are ease of computation, familiarity, and adaptability; this means that σ_y and σ_z can be altered to better fit new observations without fundamentally changing the model. Some of the limitations of Gaussian diffusion models that he disccusses are the following: they do not account for wind direction shear, they cannot describe near-calm situations, and they do not represent diffusion during stable conditions very well, when vertical diffusion may be controlled by source properties (for buoyant plumes) and lateral diffusion is due mostly to wind direction shear and meandering. Gaussian models have also been extrapolated to long range diffusion predictions over curved trajectories and changing meteorological conditions, which has no basis.

To a considerable extent the shortcomings of Gaussian models can be overcome by adjustment of σ_y and σ_z values to account for complicating factors which distort plumes from this ideal. At least this is usually possible for making practical estimates of surface concentrations. Lateral surface concentration distributions can almost always be fit reasonably well with a Gaussian curve. However, manipulations of σ_z cannot give the correct peak value of $\int \chi dy$ at the surface if an elevated plume descends, which seems to be the case for passive plumes in CBLs; in such cases, further model manipulations are required, such as forcing the mode of the Gaussian distribution to descend.

EMPIRICAL DIFFUSION COEFFICIENTS AND STABILITY TYPING

The state-of-the-art of diffusion modeling as practiced still leans heavily on methodologies developed 25 to 30 years ago. These methods are all based on fitting observations of dispersion to Gaussian formulations to determine lateral and vertical plume widths, or "diffusion coefficients", as a function of distance; these are called σ_y and σ_z, respectively. In each case, the observations were separated into categories according to some notion of stability typing, and a σ_y and σ_z curve were fit to the data in each "stability category". The stability typing schemes were different in each method, and so were the observations used, which in all cases were quite limited. It is not surprising that the derived σ_y and σ_z curves are quite different from each other. What is surprising is the degree to which these curves have been extrapolated beyond their original data base and the ways in which they have been applied uncritically to sources very different from their data base sources.

It is worthwhile to briefly review the geneses of these methods and to examine what, if any, physical bases they may have. After all, they are certain to serve as metersticks for progress in developing improved models. A fairly comprehensive discussion of existing sigma curves and stability typing schemes, including some that will be skipped here, has been given by Weber (1976). The discussion here is limited to four generic schemes. Gifford (1975, 1976) also offers an overview and considers schemes for exceptional flows, including near calm, very stable conditions, urban diffusion, diffusion over water, diffusion downwind of buildings, and diffusion in rugged terrain.

Brookhaven Curves and σ_a Typing

Stability categorization methods based on wind direction fluctuations go back at least 50 years (Giblett et al., 1932). The meteorology group at Brookhaven National Laboratory (BNL) developed a system of five stability categories based on inspection of one-hour segments of strip chart recordings of lateral wind direction fluctuations (Singer and Smith, 1966); this is a quick way to estimate σ_a. They related these categories to measurements of tracers released over about 1-hr periods at the top of a 108-m tower at BNL, which is in rolling, forested terrain of relatively large roughness. The tracers were uranine dye, oil fog, and argon 41. The lateral plume dispersion, σ_y, was measured at the surface directly, but σ_z was inferred indirectly by assuming conservation of the tracer fluxes and a Gaussian plume concentration distribution centered at the release height in the vertical (these were shown in Sec. 3 to be somewhat risky assumptions). It was also assumed that σ_y and σ_z have the same power law dependences on x, and optimum fit power laws were determined for each category.

This typing scheme has a very direct physical basis for expecting good correlation with lateral diffusion because near the source passive material goes as the wind blows, i.e., it tends to travel in straight lines at first, so that the wind direction distribution also describes the concentration distribution. In terms of Taylor's (1921) statistical theory, $\sigma_y \simeq \sigma_a x$ at small x. As material travels far enough to be influenced by a number of turbulent eddies of various scales, $\sigma_y/(\sigma_a x)$ tends to get smaller, but this dropoff is gradual and the rate of dropoff is not very different in different stabilities. Hence, σ_a or the wind direction tracer measured at the height of release over the desired sampling period correlates well with σ_y at any distance, provided that the stability has not undergone large changes since the time of release.

To address vertical diffusion, statistical theory gives us $\sigma_z \simeq \sigma_e x$ near an elevated source, where σ_e is the standard deviation of vertical wind angle. This measurement is somewhat more difficult to obtain than σ_a and is often not made. However, σ_a can substitute for σ_e as a measure of vertical diffusion if atmospheric turbulence is nearly isotropic, so that σ_y and σ_z are of the same order. This is true in NBLs and CBLs but not in SBLs, because there horizontal motion can far exceed the scale of vertical motions, which are restricted by vertical stability. Cramer (1957) noted that σ_a correlated well with σ_y measurements made day and night and with daytime values of σ_z, but the correlation with nighttime σ_z was nil. Briggs (1985a) argued on the basis of convective scaling theory that σ_a should also correlate well with vertical diffusion from a ground source while it is in the "free convective regime", with

$\sigma_z \propto H^{*1/2}t^{3/2}$. This follows from the approximate relationships $t \approx x/u$ and $\sigma_a \approx \sigma_v/u \approx 0.6w^*/u$; substituting the definition $w^* = (H^*z_i)^{1/3}$, the free convective $\sigma_z \propto (\sigma_a x)^{3/2}z_i^{-1/2}$. Because $\sigma_a^{3/2}$ ranges far more widely than $z_i^{-1/2}$, the correlation between σ_z and σ_a is strong.

Pasquill/Gifford/Turner Curves and Pasquill Typing

Pasquill's typing scheme, based on wind speed, insolation, and cloudiness, has been very widely adapted, sometimes with minor modifications. The scheme is more or less intuitive and was suggested as a way to relate diffusion rates to easily obtainable measurements. It was in use at the British Meteorological Office in 1958 and became widely used after Gifford's (1961) adaptation of the scheme for Gaussian plume modeling. Turner (1961) elaborated on the scheme somewhat by replacing subjective solar insolation categories with solar elevation and cloud cover estimates. Smith (1972) proposed a scheme using wind speed, insolation and cloudiness, but with a continuously variable numerical stability index. His σ_z curves were based on an eddy diffusivity model and included corrections for surface roughness; σ_y curves were not included.

Some confusion still exists about the origin of Pasquill's σ_y and σ_z curves, which are also the basis of Gifford's and Turner's curves, in spite of published explanations of their basis (see, for instance, Pasquill 1976a or Weber 1976). The σ_y values apply to 3-min sampling times, but frequently are misapplied to 1-hr averages with no correction. (One school of thought holds that 3-min σ_y values might give reasonable estimates of peak 1-hr concentrations occurring during the worst hours of the year; this approach would tend to overestimate typical 1-hr concentrations, however.) They are based primarily on application of the statistical theory of Hay and Pasquill (1959) to wind direction fluctuations observed at a 16 m height over relatively smooth terrain with z_0 = 3 cm. There were some very limited tracer measurements supporting the σ_y values in the range of 10-100 km, and there was much interpolation and extrapolation of curves into distances or stability categories having no data. The "neutral" (category "D") curve for σ_z was based primarily on Calder's (1949) eddy diffusivity theory applied to a surface of roughness z_0 = 3 cm. The influence of stability on σ_z at $x < 1$ km was based on results of the Prairie Grass experiments, which used near-surface releases of SO_2 (Barad, 1958). For x ranging from 1 to 100 km, some measurements of vertical gustiness in the mixed layer were used to estimate daytime σ_z, while the stable σ_z curves were more speculative. Around 1960, information on turbulence in the ABL was very fragmentary and dispersion data were quite limited in terms of range, height of release, and surface characteristics. The "ad hoc" nature of Pasquill's curves under those circumstances could be expected.

A strength of Pasquill's stability categories is that they relate strongly to the Obukhov length, L. Surface layer similarity theory predicts that L characterizes vertical turbulent motions near the surface, and it has been demonstrated by many authors that σ_z from surface releases is highly correlated with L (see, for instance, Weber et al.

1977 or Briggs, 1982b); this correlation also applies to the Richardson numbers, Ri and R_b, since they relate strongly and monotonically to L. Golder (1972) used five sets of meteorological data to develop nomograms relating R_i and L to Pasquill categories and surface roughness. Briggs (1982a) showed that L relates well to "psuedo" L's defined by $L_s = -\overline{u}^3/R_s^*$ and $L_n = -\overline{u}^3/R_n^*$, where R_s^* and R_n^* are $g/(c_p \rho T)$ times insolation and net radiation, respectively. It was pointed out that L_s and L_n simply amount to quantified forms of Pasquill's categories. The moistness of the surface affects the ratios H^*/R_s^* and H^*/R_n^*, so also affects the relationship between Pasquill categories, L_s, or L_n and L; however, this is secondary compared to the role of $\overline{u}^3$ in establishing this relationship.

Pasquill's categorization method is less effective for σ_y because scales other than the height above the ground can have more influence on lateral turbulent motions. In convective conditions, z_i affects lateral diffusion even near the ground, since $\sigma_v \simeq 0.6\ w^*$. At small x, $\sigma_y/x \simeq \sigma_a \simeq \sigma_v/\overline{u} \propto (u^*/\overline{u})|z_i/L|^{1/3}$, and z_i has as much influence on σ_y as L. In stable conditions, large horizontal meanders are very terrain sensitive and are not controlled at all by L, which is the scale limiting vertical turbulent motions. However, the Pasquill's categorization method has validity for 3-min sampling times, as represented by Pasquill's σ_y curves, since large eddies have little effect at this time scale; the small eddies are more isotropic, even in stable conditions.

A very dubious, but frequent, practice is the application of Pasquill/Gifford/Turner (P/G/T) σ_z values to pollutants from tall stacks. This was a major point of criticism in recent rural model reviews by a committee of atmospheric diffusion specialists (Fox et al., 1983). They felt that the system is based too much on surface data and that the classification system has a strong bias towards the neutral stability category. We now know that, in convective conditions, the locus of maximum concentrations from near-surface sources lift off the ground at $x \simeq 0.5\ (\overline{u}/w^*)z_i$, and thereafter the surface concentrations reduce very rapidly; this has a strong bearing on the interpretation of Pasquill's σ_z curves for "A" and "B" conditions (Briggs, 1985a). These curves are based on observed surface concentrations from Prairie Grass surface releases, and they depend on the assumption that the vertical concentration distribution is Gaussian with its maximum at the surface. Combined with the conservative tracer assumption, this gives $\sigma_z \propto (\int \chi dy)^{-1}$. However, if the locus of maximum χ elevates, causing $\int \chi dy$ to reduce rapidly at the surface, the above assumptions lead to an exaggerated growth in the calculated σ_z. Reanalyses of the Prairie Grass data using convective scaling suggest that this is what happened, since the surface concentrations, at least, behaved very much like those in laboratory tank experiments (Briggs and McDonald, 1978; Nieuwstadt, 1980; Briggs, 1985a). The Pasquill "A" σ_z curve in x expands much faster than "very unstable" curves derived from elevated sources, such as the BNL curves. Little harm is done if the P/G/T curves are applied to low sources using a Gaussian model, since that is how they were derived, but large distortions can result if they are applied to very elevated sources. For instance at about 1.8 km in "A" conditions, a plume is predicted to be four times as high as it is wide. This seems counterintuitive given the current understanding of diffusion under convective conditions (Lamb, 1979).

Högström's Sigmas and $(\partial\theta/\partial z)/U_f^2$ Typing

Remarkably little attention has been given in diffusion literature to Högström's experimental study of diffusion from elevated sources made two decades ago (Högström, 1964). The data are very useful and his analysis was detailed, scientific, and full of significant conclusions, especially regarding diffusion from elevated sources in stable conditions.

The experiments were unique in that lateral and vertical diffusion were measured optically using photographs of puffs shot from near the release point, rather than from the usual lateral camera position. Each puff consisted of a 30 s oil fog release. The puffs were assumed to advect at the rate of the mean wind speed measured at source height. Total diffusion was measured by releasing a succession of puffs over about 1 h, each one released as the former one faded. For each experiment, the standard deviations of the puff centerlines around their mean position, σ_{yc} and σ_{zc}, were determined from the photographs. In addition, the mean relative diffusion around each centerline, σ_{yr} and σ_{zr} was estimated by assuming a bi-Gaussian χ distribution and that the visible edge of the puff represents a constant value of $\int\chi dx$ (Roberts, 1923); cases of uneven or variable background were omitted. Total diffusion for each experiment was calculated by assuming vectorial addition of centerline and relative diffusion, e.g., $\sigma_y^2 = \sigma_{yc}^2 + \sigma_{yr}^2$. Altogether, 111 experiments were performed at Agesta with z_s = 50 m and at Studsvik with z_s = 24 m and 87 m (both sites are in Sweden). Conditions were mostly neutral to very stable, with only a few experiments made in slightly unstable conditions (plumes disappear more rapidly in unstable conditions).

Five different stability indices were tested for their ability to order data consisting of the most stability-dependent measure, $\sigma_{yr} \cdot \sigma_{zr}$ at x = 2000 m, for the largest group of experiments, Studsvik at 87 m. Plots were made for each index and a line of best fit was drawn by eye; correlation coefficients were computed using this function. The best correlation, 0.95, was obtained with $s = 10^5(m^3/C°\text{-}s^2)(\partial\theta/\partial z)/U_f^2$, where U_f is the speed of the "free wind" determined from routine, twice-daily rawinsondes released nearly 100 km away from this site averaged through a layer near 500 m. U_f actually worked better than u measured at at the site itself. The lowest correlations were with $\partial\overline{u}/\partial z$ instead of $\overline{u}$, and with $\partial\theta/\partial z$ alone. In all cases, $\partial\theta/\partial z$ was determined only from the 30-m and 122-m levels of the meteorological mast, so was the same as $\Delta\theta/(92m)$. Thus, Hogstrom's stability index is a form of the Bulk Richardson number, with $s \propto \Delta\theta/U_f^2$; however, it represents stability near the height of release better than surface layer stability, so is not directly related to L or Pasquill categories.

The components of relative diffusion for the Studsvik 87 m releases were highly correlated with s. At distances between 1 and 4 km, σ_{zr} approximated its neutral-condition value divided by (1 + 0.022s). Similarly,

σ_{yr} at 1 km approximated its neutral value divided by (1 + 0.012s). A very significant conclusion about σ_{yc}, which can be regarded as the "meandering" component of diffusion, is that it showed no dependence on stability. It grew roughly linearly with distance in the range 1/4 to 4 km. Another interesting conclusion made by Högström is that the wind direction shear effect on σ_{yc} was also almost independent of stability; the larger shear with increasing stability was compensated by smaller σ_z. (He estimated this effects contibution to σ_{yc} as 0.55x times σ_z times the wind direction shear between 36 and 122 m.

Högström's determinations of total σ_y for 1 h sampling times resulted in curves resembling Pasquill's, except they were about 30% smaller. The fact that Högström's σ_y values are smaller in spite of the longer averaging time might be attributed to the elevatedness of the source. At Agesta, the experimental curves for total σ_z versus x all paralleled the neutral curve, which was about 70% larger than Pasquill's neutral curve at $x = 0.1$ km but converged with it at $x \approx 5$ km. The larger initial diffusion rate at Agesta could be due to the much larger roughness ($z_0 \approx 60$ cm) compared to that represented by Pasquill's curves ($z_0 \approx 3$ cm). The stability effect on σ_z was approximated for stable cases by dividing the neutral value by (1 + 0.019s), the same type of function that described stability effects on relative diffusion. Högström (1964) developed models for both stability and terrain effects on σ_z (σ_{zc} was markedly affected by wind direction at the Studsvik site, which was only 200 m from the Baltic Sea).

It is of interest to note the relative importance of σ_{zr} and σ_{zc}, since this was one of the few studies that addresses both diffusion relative to the centerline and the diffusion of centerlines. Beyond $x = 0.5$ km, $\sigma_{zr}/\sigma_{zc} \simeq 1$ in near-neutral conditions and 1/3 to 1/2 in very stable conditions. Högström's painstaking experiments of 1960-1963 and his paper published in 1964 still stand close to the state-of-the-art because they contain unique measurements and considerable analysis, especially regarding relative diffusion.

TVA Curves and $\Delta T/\Delta z$ Typing

The TVA σ_y and σ_z curves (Carpenter et al., 1971) were developed from many years of observations of very elevated power plant plumes from a number of stacks between 75 and 150 m high (Weber, 1976). Because the plumes were very buoyant, plume rise carried them to several hundred meters. The plumes were transected both laterally and vertically by helicopters measuring their SO_2 cross sections out to distances as large as 32 km. There was no attempt to combine these profiles to form ensemble averages - probably the sampling pattern could not be repeated fast enough - so the curves represent quasi-instantaneous, or relative, diffusion. Thus, they are not adequate for predicting one-hour averages. Nonetheless the curves are instructive. The slopes of the TVA log σ_y versus log x curves range between 1/2 and 3/4 for the 1- to 32-km range of the measurements. If one considers the lateral growth of a buoyant plume due to plume rise, and assumes that $4.6\ \sigma_y \simeq$ plume width $\simeq$ plume depth $\simeq \Delta h = 1.6\ F^{1/3}\ \overline{u}^{-1}\ x^{2/3}$, the "two-thirds law", the TVA "neutral" values for σ_y are roughly approximated if the buoyancy flux $F_b = 300\ m^4/s^3$ (typical for these sources) and $u = 6$ m/sec. Although ambient turbulence or stable stratification are bound to be important in this distance range, it seems likely that growth induced by buoyant rise is also a significant component of the TVA σ_y values. The same can be said of the TVA "neutral" σ_z curve. However, the "very stable" σ_z values observed by TVA are very much supressed; with slopes of log σ_z versus log x are less than 0.2. This is in contrast to σ_z growth observed for plumes nearer to the surface, and probably reflects the absence or near absence of turbulent mixing at several hundred meters in very stable conditions. The σ_z values observed for "isothermal" to "strong inversion" cases are, in fact, about the plume thickness that would be expected to develop due to buoyant plume rise limited by stable stratification.

The stability classification system employed was based on $\Delta T/\Delta z$, the temperature gradient, measured through the plume height by helicopter. This has been criticized as inadequate to characterize turbulence, since wind speed is a stronger influence (Hanna et al., 1977). Also, as Weber (1976) points out, mean temperature gradients are adiabatic or even slightly stable above the surface layer in convective conditions; therefore, since many of the TVA observations were in the daytime and no superadiabatic temperature gradients were measured at plume heights, convective cases must be mixed in with the cases making up the "neutral" and perhaps even the "slightly stable" curves. Near the ground, $\Delta T/\Delta z$ is more responsive to heat flux and at least can distinguish between stable and unstable conditions. However, within unstable categories

$\Delta T/\Delta z$ tends to be a function of heat flux only, while $\overline{u}$ has much stronger influence on dispersion; thus, no correlation at all is seen between $\Delta T/\Delta z$ and measured σ_z or $\int \chi dy$ in unstable conditions (Weber et al., 1977; Briggs and McDonald, 1978). During stable conditions, $\Delta T/\Delta z$, Weber et al. noted that u*, H*, and L are all well correlated with each other in the surface layer and $\Delta T/\Delta z$ correlates fairly well with near-surface vertical diffusion, provided it is measured through a substantial layer roughly coinciding with the plume layer. (The best correlations with σ_z were obtained with $(z_2 - z_1) > 5$m and $z_2/z_1 > 2$, where z_2 and z_1 are the upper and lower heights of temperature measurement, respectively).

EVALUATIONS OF SIGMA CURVES AND TYPING SCHEMES

We find surprisingly few comparisons of diffusion data with rival schemes for σ_y, σ_z, and stability classification. There have been dozens of papers demonstrating that different stability schemes disagree with each other, but only a few papers address the question of which one is best. Weil (1977) compared the performance of different Gaussian model schemes applied to three power plants, which involves the additional uncertainties associated with buoyant plume rise. The Brookhaven σ_y and σ_z values worked best, but they were chosen by an algorithm of Weil's own devising based on wind speed and $\Delta\theta$ (this most closely relates to bulk Richardson number classification). The TVA scheme was the poorest performer. Pasquill's scheme was intermediate in performance, even though its basis was in surface source data; its performance was greatly improved by displacing the stability category towards the unstable direction by one category. Similar results were obtained at all three power plants, except for a few exceptional cases at Chalk Point; this was the only plant having σ_a measurements, so it seems that σ_a classification schemes were not adequately tested.

Quantitative tests of the correlation between various stability indices and vertical diffusion from surface sources were made by Weber et al., (1977). Only linear correlations were tried, and σ_z was mostly inferred from surface-arc $\int \chi dy$ measurements and the Gaussian plume assumption, using the Prairie Grass, Green Glow, and NRTS data sets. The indices tested were ΔT, net radiation, σ_a, σ_e, $\overline{u}$, the gradient Richardson number (Ri), and z/L. For unstable cases, ΔT and net radiation did not correlate with σ_z at all. The wind direction variances had mixed results with mostly rather weak correlations, but the wind vanes used in these early experiments may have been poor. The best correlations were with Ri and z/L, which are closely related to each other. For stable conditions, σ_e gave the most consistently good correlations with σ_z, but Ri, z/L, and ΔT also gave good correlations. At the farthest distance, 3.2 km, all correlations fell to 0.5 or less. The Prairie Grass data set provided the best correlations; these data were further used by Briggs and McDonald (1978), who compared $(\int \chi dy)^{-1}$ with various indices and nondimensionalizations. The single best stability index was L, but $\overline{u}^2/\Delta\theta$ worked almost as well and it is much easier to measure; $\overline{u^2}$, by itself, applied separately to stable and unstable data categories, worked slightly less well but still provided good correlation with $\int \chi dy$; this supports Pasquill's scheme for ground-level sources, since it depends most strongly on $\overline{u}$. Measured between 1 and 8 m, $\Delta\theta$ also provided good correlation in stable conditions, but none at all in unstable conditions. Good ordering of the data was obtained with nondimensional plots involving L, $\overline{u}^2/\Delta\theta$, or, for unstable cases, convective scaling with z_i and $w*$; L-scaling had a slight advantage in stable conditions, as did convective scaling in unstable conditions.

Comparisons of σ_y and σ_z schemes against data from a large number of both surface- and elevated-source experiments were made by Irwin (1983). He tested the P/G/T scheme, with classification by Turner's (1970) method, along with five schemes based on wind direction fluctuations, with $\sigma_y/(\sigma_a x)$ or $\sigma_z/(\sigma_e x)$ equal to f(x) or f(t), where $t = x/\overline{u}$. The functions f(x) and f(t) are all unity at very small x, consistent with statistical theory. The P/G/T classification scheme could be applied in a consistent way only to a very limited data set, the Prairie Grass and Karlsruhe experiments. The P/G/T system worked reasonably well, almost as well as the best σ_a-σ_e models, in stable conditions (there were adequate data for surface sources only). However, it consistently underestimated σ_y in unstable conditions, which is not surprising considering that it represents only 3-min averages for σ_y; for the 10-min surface releases of Prairie Grass, a factor of 1.6 increase in σ_y would produce better agreement with observations, and for the 20- to 60-min elevated releases at Karlsruhe, a factor of 2.7 increase is needed. The P/G/T system also greatly underestimated σ_z in unstable conditions inferred from the Karlsruhe experiments, by about a factor of 3. While its performance for the Prairie Grass observations was much better, we must remember that Pasquill's σ_z curves were largely based on the Prairie Grass data. All the σ_a-σ_e schemes were tested on the full data set. They all performed fairly well for σ_y estimates, except that Pasquill's (1976a) f(x) dropped too rapidly at large x in stable conditions. Cramer's (1976) f(x) did not decrease fast enough for the surface releases, but his model was designed for elevated releases and was the superior performer for elevated source σ_y. For σ_z, Cramer's model, with $f(x) = 1$ at all x, performed slightly better than others in unstable conditions (the full model was not tested for stable conditions because it requires specification of h). Draxler's (1976) model and a simplified form of its, proposed by Irwin, gave the best results overall for σ_z, these models were substantially outperformed only in the case of elevated-source σ_y in stable conditions, by Cramer's model.

Many of the field experiments used in the above evaluations suffered from deposition losses and less reliable σ_a and σ_e measurements than given by newer instruments. The recent field study in Copenhagen using an elevated source of SF_6 should be more dependable and has been used to test published Gaussian model schemes (Gryning and Lyck, 1984). For σ_y, the methods using σ_a produced notably less scatter in predicted-to-observed ratios than the methods based on Pasquill categories. Hay and Pasquill's (1959) statistical theory technique and Draxler's (1976) scheme clearly gave the best results, with no significant bias. For predictions of surface-measured $\int \chi dy$ using the Gaussian model with reflection at $z = z_i$, the seven models for σ_z gave roughly the same amount of scatter. This suggests that most of the scatter is due to the stochastic variability of the atmosphere itself; for these 60-min averages in neutral to moderately unstable conditions, then, 50% of the $\int \chi dy$ fell within about a factor of 1.2 of the mean and 75% of the observations fell within about a factor of 1.5 of the mean prediction; even the "perfect model" may produce little improvement in this scatter. Three models had insignificant bias: Hay and Pasquill (1959), Draxler (1976), and the P/G/T method (Turner, 1970). The σ_z enhancement due to roughness and urban-enhanced heat flux recommended by Pasquill (1976a) resulted in underestimates of $\int \chi dy$ for this suburban area about a factor of 1.4. The largest mean underestimate, of about a factor of 2, was by a model taken from Hanna et al. (1982); although Gryning and Lyck did not state the model specifically, it probably is the Briggs, "urban" σ_y and σ_x curves, which are based on curves from the St. Louis diffusion experiment (McElroy and Pooler, 1968). This suggests that urban σ_y and σ_z curves give too fast a growth rate for suburban locations. For ratios of predicted-to-observed maximum concentrations for Copenhagen, all four methods using σ_a to predict σ_y gave the least scatter; three of these also gave insignificant bias: Hay and Pasquill (1959), Draxler (1976), and Pasquill (1976). For overall performance in predicting σ_y, $\int \chi dy$, and χ_{max}, it appears that the Hay and Pasquill and the Draxler methods were clear winners in this contest.

To summarize these evaluations, it appears that models based on $\sigma_y = \sigma_a x \, F(t)$, with a moderate degree of sophistication in the specification of f(t), like Draxler (1976), give the best results so far for lateral dispersion. The Draxler (1976) formula for σ_z, based on σ_e measurements, also appears to be the most dependable method for predicting vertical dispersion, although several similar methods are competitive with it. This supports the recommendations of scientists at the various workshops since 1977 that favor characterization of σ_y and σ_z with σ_a and σ_e. The P/G/T schemes, based primarily on categorization by wind speed and secondarily on insolation or cloudiness, have an inconsistent performance record. For σ_y, the Turner (1970) method works about as well as σ_a methods for surface releases in stable conditions at a smooth, flat site (Prairie Grass), but this is probably not true in more typical terrain, as large, slow meanders are likely to greatly increase σ_y and σ_a. This method considerably underestimates 1-hr average σ_y values in unstable conditions. For σ_z, it tends to give too rapid growth for elevated sources, but remains competitive with the σ_e methods for the surface releases of the Prairie Grass data set (which is the main basis for the P/G/T σ_z curves). Categorization schemes based on ΔT or $\Delta\theta$ alone are definitely out of favor, although the Briggs and McDonald (1978) results indicate usefulness for surface-release σ_z in stable conditions only (σ_e was the superior index, however). For vertical diffusion from a surface release, the Obukhov length, L, was the most effective stability index, followed by its simple cousin, $\overline{u}^2/\Delta\theta$. Even $\overline{u}^2$ alone, with data segregated into stable and unstable categories, proved a worthy index for diffusion rates, far better than $\Delta\theta$ alone.

SECTION 5

DIFFUSION MODELING ALTERNATIVES

The "conventional" models discussed in Section 4 all assume a Gaussian plume and all are mostly empirical. However, we find that most of the stability classification schemes have some physical basis. The Brookhaven σ_a classification relates directly to σ_y at short distances, with the relationship degrading very slowly with distance. It relates well to σ_z when conditions are nearly isotropic, i.e., lateral and vertical turbulence velocities and length scales are about the same; this condition holds best for elevated sources in neutal and unstable conditions. Pasquill's classification based on $\overline{u}$ and insolation or cloudiness is crudely related to the Obukhov length, L, the most important length scale affecting vertical turbulence in the lower part of the ABL. Thus, it correlates best with σ_z from surface or low sources. It correlates fairly well with short-term averages (3 min) of σ_y, since the small eddies affecting this time scale are more isotropic, but considerably underestimates typical 1-hr averages of σ_y. Högström's ($\partial\theta/\partial z/U_f^2$) classification is a bulk Richardson number, which is strongly related to L if $\partial\theta/\partial z$ is measured near the surface. Measured near 100 m, it is a good index of the relative vertical stability at heights of elevated releases. For neutral to stable conditions it correlated very well with σ_z, both relative and total, and with σ_{yr}, the relative part of lateral diffusion (this was accomplished by smaller eddies which are more isotropic); however, it did not correlate with the total (time-averaged)

σ_y. TVA's index $\Delta T/\Delta z$ and variations on it like ΔT, $\Delta\theta$, or $\partial\theta/\partial z$ are very weak indicators of the atmosphere's diffusive capacity, because wind speed has a stronger effect than the static stability.

Thus, these classification schemes, except for $\Delta T/\Delta z$, each have some areas of success as diffusion indicators, and none is "universal". The Gaussian plume assumption has been a useful workhorse, but we have identified some diffusion conditions during which it leads us astray, as the atmosphere takes on more complex behavior. The worst aspect of the models reviewed in the previous section is that each was built on a very small set of diffusion measurements, measurements that had quite a few shortcomings; since the curves are empirical, there is simply no basis for extrapolating them to much larger x or using them for all types of sources and terrains. Here is where more scientific models become practical tools. If they can be validated for enough parameter combinations within their range of applicability, there is some basis for confidence in extrapolations and interpolations.

Over homogeneous terrain and under near steady conditions, typical Gaussian models predict peak concentrations within about a factor of 2. There is widespread belief that this kind of performance can be improved upon by models which incorporate adequate physics, if they can be supported by adequate meteorological measurements. This section describes six general methods of approach that have had proven success in some areas of testing; none of them appear to be the universal solution for all diffusion problems, but a combination of approaches should be able to provide the large step in modeling improvement that is greatly needed.

SURFACE LAYER SIMILARITY MODELS

In the early 1960's, there were a number of attempts to rationalize atmospheric dispersion rates in terms of fundamental boundary layer parameters that were discussed in Sec. 2. The inspiration for this work was the so-called "similarity theory" of Monin and Obukhov (1954), which was becoming a successful tool for describing profiles in the surface layer of wind speed, temperature, humidity, and vertical turbulence quantities. This theory assumes that vertical turbulence near the surface is a function of height and velocity and length scales based on surface fluxes of heat (or buoyancy) and momentum (these scales were defined in Sec. 2). Thus σ_w is assumed to be $u*$ times a function of normalized height, z/L, on dimensional grounds. Vertical gradients of $\overline{u}$, T, water vapor, etc., are assumed equal to their fluxes at the surface divided by $(u*z)$, times some function of z/L. These functions must be determined empirically and may be different for different quantities. The usefulness of this approach, essentially a dimensional analysis, was shown decisively by the results of the 1968 Kansas field experiment (Businger et al., 1971; Kaimal et al., 1972).

If the theory describes the structure of turbulence near the ground, then it should also be capable of describing turbulent diffusion of passive tracers released at the surface; this was tried in 1957 by Kazansky and Monin for describing the shape of smoke plumes (see Monin, 1959). Gifford (1962) reviewed progress with similarity theory up to that time and developed predictions of diffusion from the surface starting with the equation

$$d\overline{z}/dt = cu_*\phi(\overline{z}/L), \qquad (5.1)$$

where $\overline{z}$ is the mean height of the concentration, c is a dimensionless constant, and ϕ is a universal function of z/L. Other early examples of the application of Monin-Obukhov similarity theory to diffusion are the papers by Cermak (1963), which used mostly wind tunnel diffusion observations, and by Panofsky and Prasad (1965). The latter authors noted that lateral dispersion does not fit the theory as well as vertical dispersion; this observation has been reconfirmed by many others since then, and seems to follow from the fact that lateral turbulence is influenced by z_i at day and by topographically influenced meanders at night. Thus, surface layer similarity theory omits significant factors affecting lateral dispersion.

The subject lay in a relatively dormant state, stymied in part by the difficulty of measuring L, until after analyses of the 1968 Kansas wind and temperature profile data were published. Chaudhry and Meroney (1973) made use of the Businger et al., (1971) relationship between Richardson number, $R_i = (g/\theta)(\partial\theta/\partial z)/(\partial\overline{u}/\partial z)^2$, and L, to make improved estimates of L for the Prairie Grass data. They also advanced the theoretical argument that $\phi = \phi_h^{-1}$, where ϕ_h is the dimensionless temperature gradient given by $ku_*\partial\overline{\theta}/\partial z = (\overline{w'\theta'}/z)\phi_h$, and θ is potential temperature. The argument was made using an eddy diffusivity approach. The simplest way to do this is to recast Gifford's equation as

$$d\overline{z}^2/dt = 2bu_*\overline{z}\phi = 2K \,, \qquad (5.2)$$

where K is an eddy diffusivity for passive material diffusing from the surface. Chaudry and Meroney argued that this diffusion mechanism is similar to that for heat flux from the surface, so $K \propto K_h$, the eddy diffusivity for heat; this is defined by $\overline{w'\theta'} = K_h \partial\overline{\theta}/\partial z$. Substitution of the profile law gives $K_h = k\overline{u}_* z/\phi_h$ and, therefore, $\phi = \phi_h^{-1}$. These authors were able to demonstrate that this hypothesis leads to a much better fit with the Prairie Grass data than does Monin's (1959) hypothesis for ϕ, which is too weakly dependent on L. They used the observations at x = 100 m, the only arc having samplers on towers; this permitted direct determination of $\overline{z}$ (no Gaussian assumption required).

Horst (1979) carried this analysis further using all the arc distances of the Prairie Grass experiment, ranging from 50 to 800 m, but he had to assume a vertical distribution function to relate $\overline{z}$ to $\int \chi dy$ at the surface, since only near-surface samplers were used. Based on Prairie Grass profiles at x = 100 m and other evidence, he chose a modified Gaussian profile assumption: $\chi^{-1} \propto \exp(z/b\overline{z})^r$ with r = 1.5. He then showed that $K = K_h$ leads to a satisfactory fit with observations, much better than $K = K_m$, which had been suggested by Pasquill. He also compared the prediction with $\int \chi dy$ observed at x = 3.2 km at the National Reactor Testing Station (Islitzer and Dumbauld, 1963), finding more scatter but definite skill using $K = K_h$. Van Ulden (1978) also used $K = K_h$ in a similar analysis for arc distances x = 50, 200, and 800 m; further, he made an allowance for the non-zero height of the Prairie Grass samplers, z = 1.5 m. In a recent re-analysis of the Prairie Grass data, Gryning, van Ulden and Larsen (1983) again demonstrate a good fit of the observations using surface layer similarity and $K = K_h$, with poorer results using K_m. They find improved fits when they include an estimate of deposition in their model, as SO_2 is now known to deposit on vegetation.

Two recent papers attempt to simplify application of surface layer similarity to vertical diffusion from the surface. Both of these utilize the Prairie Grass data. Venkatram (1982) argues that both the neutral and very stable asymptotes for $\int \chi dy$ at the surface are independent of $\overline{u}$. (To simplify discussion, $\overline{z}$ will be used in place of σ_z and cloud height h, since they are all closely related.) For surface, concentrations from surfaces sources $\int \chi dy \propto Q/(\overline{u}\,\overline{z})$. As $\overline{z} \propto u^*t \propto (u^*/u)x$ for the neutral case, the $\overline{u}$ dependence cancels out: $\int \chi dy \propto Q/(u^*x)$. For extreme stability, with $\overline{z} >> L$, Venkatram assumed $\overline{u} \propto u^*z/L$ and a constant eddy diffusivity, $K \propto u^*L$. Substituting this into $\overline{z} \propto (Kt)^{1/2}$, with $t \propto x/\overline{u}$ and the above $\overline{u}$, we get $\overline{z} \propto L^{2/3}x^{1/3}$ and $\int \chi dy \propto Q/(u^*L^{1/3}x^{2/3})$. (The existence of this asymptote in the ABL can be questioned in light of new knowledge about SBLs, because h/L never gets large enough.) Venkatram used profile determinations of L and u* to nondimensionalize the data, showing that the two asymptotes fit the neutral-stable Prairie Grass data rather well, except for some widely scattered points at larger x/L values. Briggs (1982b) derived the same asymptotic relationships, arguing that there is a very slight dependence on $\overline{z}/z_0$ at intermediate values of x/L, which can be neglected. He also extended this approach, using cruder approximations, into the convective regime, and suggested analytical expressions for $Q/(u^*x \int \chi dy)$ versus x/L. The numerical constants in his analysis differ from Venkatram's asymptotes by about 20%, partly because Briggs adjusted for the non-equality of the source and receptor heights in the experiment. Venkatram (1980) also demonstrated a strong correspondence between u*, L and $\overline{u}$ in stable conditions, which could be used to simplify application of these results.

It can be concluded that surface layer similarity models can give very good results for vertical diffusion from near-surface sources at distances up to 1 or 2 km. The theory does not contain enough information about the variation of turbulence with height to give good results for sources at more than 1/10 the ABL height. Except for averaging times of less than 10 min, for which diffusion is accomplished mostly by small, isotropic eddies, this approach can only predict the minimal (rather than typical) lateral diffusion. This is because it contains no physics relating to large horizontal eddies.

GRADIENT TRANSFER AND HIGHER ORDER CLOSURE MODELS

One of the earliest and most frequently used approaches to prediction of diffusion in the atmosphere is the gradient transfer assumption, or "eddy diffusivity" theory, or "K theory". It assumes that the local turbulent flux of a material is directly proportional to the mean local gradient of that material, and is aligned into that gradient (towards lower concentration). The constant of proportionality, K, is called the eddy diffusivity and is supposed to be a property of the flow, rather than of the substance. Thus, in a steady state situation, K should be only a function of position (x, y, z) and boundary conditions. For just the vertical component of turbulent flux of substance C, we can write the gradient transfer assumption quite simply

$$\overline{w'C'} = -K\ \partial\overline{C}/\partial z \quad . \tag{5.3}$$

If there is no mean advection, then the rate of change of concentration at each point is given by

$$\partial\overline{C}/\partial t = -\ \partial\overline{w'C'}/\partial z = \partial(K\partial C/\partial z)/\partial z. \tag{5.4}$$

This approach is tempting in its simplicity, and builds on a long history of mathematical treatments of molecular diffusion. In that context, K is just 1/3 of the mean molecular speed times the mean free path length between molecular collisions. The most direct analog for turbulent diffusion is K proportional to the mean eddy velocity times a mean eddy diameter, or turbulence length scale (such as σ_w times the Lagrangian time scale). However, this analogy is often inappropriate, and there are only certain situations in atmospheric diffusion for which K theory is approximately valid. Pasquill (1974, section 3.1) has discussed this at some length. The question of validity has to do with the difference between diffusion and dispersion, terms that usually are used as if they were interchangable. "Diffuse" in its Latin root means to pour or to spread out, and seems to connote smearing. Thus, local flux proportional to local gradient seems a good description of "diffusion". "Disperse" in its Latin root means to scatter, which better describes what big eddies do to small plumes - wave them this way and that. The relative size of the eddies and the plume or puff is the key. The idea that local mean concentration gradients have anything to do with "dispersion" of small plumes by large eddies seems absurd, if we try to visualize this process. Yet K theory has been indiscriminately applied to dispersive situations. On the other hand, it is intuitively a good assumption to apply to "diffusion" of large plumes by small eddies. Pushed to its limit, we might expect this assumption to hold as long as mean concentration gradients do not change drastically on the scale of the most effective eddies.

There are three cases of atmospheric diffusion which might be describable using ordinary K theory, as described above. The diffusion of material released near the surface is one. This application actually occupies the fuzzy region between validity and non-validity, because the most effective eddies doing the upward mixing are those of the order of the plume thickness or slightly larger; it can be viewed as an approach that might work, especially when the eddy structure maintains similarity relative to the height above the ground. This condition is most closely met in the lower part of the NBL. Here, $K \propto u^*z$ has proven validity, as shown by many of the papers referenced in the last subsection. The SBL is also a reasonable candidate for K theory, because the eddy size becomes gradually restricted by the scale L as z/L becomes larger and vertical diffusion becomes more of a diffusive process (eddies small compared to scale of variations in mean gradients). The unstable case is just the opposite; the higher the material mixes, the larger the dominate eddy size until the dispersant is drawn into thermals which lift the material all the way to z_i. Here, the eddy diffusivity assumption is inappropriate. At best, it might apply as long as the maximum mean concentration remains on the ground and the vertical distribution is quasi-Gaussian; there is growing evidence that this condition holds only for $t < 0.5\ z_i/w^*$, or $\sigma_z < 0.3z_i$. Deardorff and Willis (1975) demonstrated conclusively, using the gradient transfer equation in reverse, that K would have to be a function of time of release to explain the concentration fields they observed in their convective tank experiments. (K is supposed to be a property of the flow only, not of release time.) Furthermore, K would have to become negative to explain their results; that is, at times the mean flux of concentrations went against the gradient. This happens when thermals continue to draw material in from a region of lower mean concentration near the ground and transport it to heights of higher mean concentration.

Eddy diffusivity treatments for vertical diffusion from the surface have been developed by Smith (1972) and by Pasquill (1978), using the most current knowledge about boundary layer turbulence in the assumptions $K \propto \varepsilon^{1/3}\lambda_m{}^{4/3}$ and $K \propto \sigma_w\lambda_m$ (ε is the turbulent energy dissipation rate, σ_w is the variance of vertical velocity, and λ_m is the scale of maximum turbulent energy, determined from spectra). Smith correlated his numerical solutions with a classification system similar to Pasquill's, with a simple graphical correction for the effect of roughness length on σ_z, in an attempt to put his results in a form suitable for practical application.

The second case for which K theory might have some validity is that of vertical diffusion or an elevated release in stable conditions, but only after the plume thickness is of the order of L or Λ, the "local" Obukhov length defined by Nieuwstadt (1984a,b). The stable stratification limits vertical mixing motions to a depth comparable to L or Λ, except near the ground where z is a more limiting factor. Observations of layers of near-constant $\partial\overline{\theta}/\partial z$ and $\partial\overline{u}/\partial z$ in stable, dry boundary layers above $z \approx L$ indicate that $K_m \approx K_h \approx 0.08u^*L$, where K_m and K_h are eddy diffusivities for momentum and heat, respectively. We would expect K for the diffusion of passive matter to be about the same. Ordinary K theory has no validity for elevated plumes in neutral and unstable conditions, because the plume thickness is then smaller than the most active eddies, which are dispersive rather than diffusive.

The third possible case for limited validity for K theory is that of lateral diffusion in convective conditions at distances beyond several $(\overline{u}/w^*)z_i$. By this time the plume width is several z_i, larger than the dominant cell size (about 1.5 z_i). The asymptote for σ_y suggested by Deardorff and Willis (1975), based on their laboratory tank experiments, is $0.53\ (w^*z_it)^{1/2}$, which would result from $K = 0.14\ w^*z_i$ (note that, in this case, convective scaling with w^* and z_i applies, rather than surface layer scaling with u^* and L). However, in the atmosphere, where wind shear plays a role and there are no horizontal restrictions like the tank walls, there are horizontal eddies or rolls of much larger scale than z_i in the line of mean wind direction. Some indication of this is given in the report on the Minnesota convective boundary layer experiment (Kaimal, et. al., 1976). Very frequently, wind direction shifts are observed to occur on time scales of about one hour on convective days (personal observation). Then continuous plumes are still "dispersing" on these time scales, and a $\sigma_y \propto x^{1/2}$ or $\propto t^{1/2}$ regime may be short-lived, about 1 hr at most. Lateral dispersion observations in neutral and unstable conditions have mostly yielded best fit power laws between $x^{0.7}$ and $x^{0.9}$, in contrast to the $x^{0.5}$ asymptote given by K theory. This suggests that plumes are usually influenced by larger and larger horizontal eddies, contrary to K theory assumptions.

One unrealistic consequence of any diffusivity model is that non-zero concentrations are predicted out to infinity, implying an infinite velocity of propagation. An example is the Gaussian distribution, which results when K is assumed constant. Although these far-flung concentrations are quite insignificant in magnitude, the physical unreality of this aspect of such models is discomforting. An alternative approach, called the telegraph equation, has been developed by Monin and Yaglom (1965). Essentially, it restricts the rate of change of flux by adding the term $\tau\ \partial\overline{w'C'}/\partial t$ to $\overline{w'C'}$ in the flux-gradient equation, where τ is a limiting time scale. The result is that there is a limited velocity of propagation proportional to $(K/\tau)^{1/2}$, and the flanks of the concentration distribution have a cutoff. Closer to the center of the distribution there is a slight change, but not enough to have any practical consequence (see Pasquill 1974, Fig. 3.2).

The most serious limitation of simple eddy diffusivity approaches is that non-local effects, such as those brought about by large turbulent eddies, are neglected. The local turbulent flux of concentration is assumed to be a function of only the local gradient of concentration. When large eddies bring in fluid containing zero concentration, which therefore comes from regions of zero gradient, this assumption of dependence only on local gradient is obviously wrong. Several attempts have been made to construct models which avoid this deficiency. One involves a more sophisticated version of K-theory called spectral diffusivity (Christensen and Prahm, 1976; Berkowicz and Prahm, 1980). The concentration field is described in terms of Fourier components, with c(k) being the amplitude of the Fourier mode of concentration with wavenumber k (waves per unit length). The eddy diffusivity is assumed to be a function of k; thus, K(k) can account for the variability of turbulent energy in each range of eddy size. The eddy diffusivity equation for a Fourier mode becomes $\partial c(k)/\partial t = -k^2 K(k) c(k)$. In a sense, this model is non-local, since each Fourier component represents concentration spanning the whole dominion under consideration. A particular form that has been proposed for K is $K_0/(1 + Bk^{4/3})$. This gives a constant K, with the same result as ordinary diffusivity theory, for very large wavelengths (small k), i.e., for variations in concentration on a scale large compared to energetic turbulent eddies. For small wavelengths $K \propto \varepsilon^{1/3} k^{-4/3}$, where ε is the turbulent energy dissipation rate; this is the asymptote for the "inertial subrange" of turbulent eddies driven by the larger eddies. Prahm, Berkowicz and Christensen (1979) applied this theory to time averaged plume concentrations, adding the effect of "meandering" of the plume centerline by means of a "spectral phase diffusivity coefficient" that is made to fit statistical theory predictions for large and small asymptotes of k and t. The resulting plume growth is, not surprisingly, in agreement with the main results of statistical theory, but the shape of a narrow plume is different from Gaussian. There is no experimental confirmation of this shape as yet, and its practical significance seems doubtful. The theory has primarily been applied to two dimensional global transport models, with eddy sizes ranging up to several thousand kilometers.

Another type of non-local K theory was suggested by Voloshchuk (1976). In this model, K is both a function of z and of a displacement distance, l. If K(z,l) has a non-zero value only for l = 0, the equation reduces to the telegraph equation for the local case, like that developed by Monin and Yaglom (1965). While this method appears to have potential for accounting for non-local concentration gradients, the actual specification of K(z,l) for various stabilities might be quite difficult, considering the effort that has been spent on trying to specify K for a single variable, z.

A new mechanism for turbulent transfer has been developed by Stull (1984) and applied in one form by Stull and Hasegawa (1984). This mechanism is called transilient turbulence theory, from the Latin verb "transilire" meaning to jump over or leap across. Stull introduces the mechanism in a finite difference framework, then goes to a continuous representation. In either form, the mechanism adjusts the concentration at a point by considering the concentrations at all neighboring points which can be influenced by the flow. For example, consider only vertical transfer, for which the concentration at a point is influenced by all of the points in the vertical. The formal expressions for the continuous case are similar to the equations for radiative transfer in an absorbing, but nonscattering medium. Fiedler (1984) independently developed a similar concept, but started from a different perspective, namely the spectral diffusivity concept of Berkowicz and Prahm (1980). Stull's approach was to visualize turbulence as consisting of a spectrum of eddy sizes each contributing to the turbulent transfer. The interesting feature of the mechanism is that the coefficients defining the transfer process can be specified ahead of time, or can be made responsive to changing flow conditions. The coefficients can be made to represent any type turbulent flow. The same model with different coefficients can represent diffusion in a CBL or in a SBL. Further, Stull points out that the transilient approach can also be used in combination with a second order closure model to close the higher moments. This new development has much promise.

K theory is considered "first-order" closure, meaning that closure assumptions are applied to the second moments of turbulent quantities, such as $\overline{w'C'}$ and $\overline{w'u'}$, that relate them to first order quantities like $\overline{C}$ and $\overline{u}$. About 10 years ago, work accelerated on "second-order" closure models for turbulent boundary layers and diffusion, carrying this approach to a new level. These models begin with the exact equations for second moment terms like concentration flux or momentum flux, which are needed to solve for $\partial\overline{C}/\partial t$ or $\partial\overline{u}/\partial t$. As in all turbulence models, the exact equations contain terms with new unknowns, and always more unknowns than equations, so closure assumptions are still necessary. In this case, the new unknowns are third moments, such as $\overline{u'w'C'}$, and assumptions must be made relating them to terms already being computed, i.e., second moments or mean quantities. These assumptions may be just as questionable as assumptions made about K in first order theory, and the third moment terms may be modeled quite crudely; the advantage is that second moment terms like turbulent fluxes do not depend on these assumptions alone, but also depend on exact terms; thus, they are less susceptible to error on account of the failure of a closure assumption. Nevertheless, the output of second-order closure models depends on the performance of the closure assumptions, and these should be scrutinized for proper physical characteristics and tested against data in a variety of flow situations. One shortcoming that second order closure models so far share with K theory is that the closures are local; thus, the effect of large eddies on the third moment terms cannot be handled properly.

One of the advantages of second-order closure modeling of diffusion is that buoyancy-driven vertical mixing is included; this is done with the term $(g/\overline{\theta})\overline{\theta'C'}$ in the equation for $\partial\overline{w'C'}/\partial t$. Whereas gradient transfer theory fails to model the lifting of the mode of maximum concentration from a surface source in convective conditions, Lewellen and Teske (1976) demonstrated a second-order closure model that does this. The qualitative features of $\int \chi dy$ that were observed by Willis and Deardorff (1976) in the laboratory are reproduced quite well, but the surface concentrations do not reduce quite as fast in this numerical model. Pasquill (1978) also reports some success with second-order models in the case of vertical diffusion from an elevated source, another case which cannot be handled by conventional K theory. Second-order closure models have also been used to study the SBL, as reported in Sec. 2 Wyngaard (1975) and Rao and Snodgrass (1978) were able to reproduce in their models the approximate behavior with time of quantities like heat flux, friction velocity, and boundary layer height, h, which is of vital relevance to diffusion modeling. It evolves rapidly after sunset, and a low or medium level source could be below or above h; this determines whether or not vertical diffusion occurs at all.

CONVECTIVE SCALING MODELS

A very active area of research at present is the application of convective scaling to daytime diffusion models. The idea was first applied to atmospheric boundary layer turbulence by Deardorff (1970b), who first tried this scaling in the context of a large eddy simulation model of neutral and unstable boundary layers; this model will be discussed in the next subsection. Subsequent successes using this scaling in the context of CBL turbulence measurements were reviewed in Sec. 2. For $|z_i/L|$ equal to 4.5 and larger, Deardorff (1972) found that $w*$ scales the magnitude of turbulent velocities better than $u*$. Typical daytime values of convective and surface layer scaling quantities are: $z_i \sim 1000$ m, $u* \sim 0.4$ m/s, and $H* \sim 0.004$ m^2/s^3. This gives $L \sim -40$ m and $|z_i/L| \sim 25$, which is quite convective according to Deardorff's results. The criterion $|z_i/L| > 4.5$ would be met even with 1/5th the above values of z_i or $H*$, or with these values unchanged and $u* = 0.7$ m/s; this typically corresponds to wind speeds of about 6 m/s near the ground; $\overline{u} = 6$ m/s with moderate insolation corresponds with stability class "C-D" in Pasquill's table, between "neutral" and "slightly unstable". We can conclude that convective scaling would be appropriate for most of the daytime on most days at most sites on land. One condition that produces sustained maximum average ground concentrations near large power plants is a sunny summer day with low wind speed; these are very convective conditions, with $|z_i/L|$ much larger than 100. This is a case which should certainly benefit from the convective scaling models recently under development.

Application of the convective scaling idea to diffusion began with the now-famous convective tank experiments of Willis and Deardorff (1974, 1976, 1978, 1981, 1983). These experiments and their controversial results were discussed in Section 3 under "Laboratory Experiments". These results were unexpected and cast much doubt on the ability of conventional Gaussian models to accurately model daytime diffusion. Naturally, there was much interest in verifying these results outside the modeling tank. The numerical experiments of Lamb, using Deardorff's (1972) model, was a first attempt. The results were qualitatively quite similar to those of the tank experiments, with nearly the same descent rate of the maximum concentrations from elevated releases, but with slightly less decrease in surface concentrations at the larger distances. Utilizing field data that were already available, both Briggs and McDonald (1978) and Nieuwstadt (1980) demonstrated that the "Prairie Grass" surface values of $\int \chi dy$ were quite similar to those for the lowest release in the tank when nondimensionalized using convective scaling. The field experiments showed more scatter; this was attributed to the short release time, only 10 minutes, which is inadequate to sample many eddies (a typical "eddy passage time" is 1.5 z_i/u, which can range from 2 to 30 minutes or more). The Prairie Grass σ_y/z_i values were about 40% larger than those in the tank; part of this may be due to the absence of horizontal motion restriction in the atmosphere.

It is ironic that field experiments done in the two decades following Prairie Grass did not include sufficient measurements to determine z_i, a parameter that turns out to be indispensible for modeling diffusion in convective conditions (the 1956 experiment included aircraft soundings of temperature and humidity, as well as rawinsondes, so z_i was easy to determine in retrospect). A recent field experiment designed to test convective scaling on diffusion in the atmosphere, especially for comparison with the tank experiments, was carried out at the Boulder Atmospheric Observatory in August and September of 1982 and 1983; this was a cooperative experiment between the U.S. EPA and NOAA's Wave Propagation Laboratory. It is described in Section 3. Some preliminary results have been reported by Moninger et. al., (1983), Eberhard et al. (1985) and by Briggs (1985a). In the first paper, a striking resemblence was seen for nondimensionalized $\int \chi dy$ between a 29-min period with $z_s/z_i = 0.45$ and the tank experiment for $z_s/z_i = 0.49$. However, it is not yet known how much the 0.3 m/s fall speed of the chaff tracer distorts the distributions (this concern motivated the simultaneous side-by-side releases with oil fog). In Briggs (1985a), five periods of chaff distribution were analyzed in comparisons with the tank and numerical experiments and three other field experiments including: the most convective cases from Project Prairie Grass, the five most convective 30-min periods of SF_6 releases at Cabauw, the Netherlands, which really were only marginally convective (van Duran and Nieuwstadt, 1980), and, finally, some SO_2 measurements at the Morgantown power plant (Weil, 1979). The latter observations were screened to remove buoyancy-dominated cases. The results of these comparisons show a good deal of consistency, in convective scaling. All experiments support the tank-numerical σ_y/z_i curves, except for two BAO periods with very large wind direction shear; these had about double the σ_y. The near-surface $z_i\bar{u}\int\chi dy/Q$ for one of the surface chaff releases dropped to a value about 30% lower than the tank minimum, but all the remaining data were in better agreement. The most significant result was for the maximum surface value of $\int\chi dy$ from elevated sources; there was a definite trend with increasing z_s/z_i, with $C_y' = z_s\bar{u}\int\chi dy/Q \approx 0.48(1 + 2z_s/z_i)$. Conventional Gaussian models give $C_y' = 0.48$, regardless of source height. These maxima were all found in the neighborhood of $x = 2\,\bar{u}\,z_s/w^*$.

Another recent confirmation of convective scaling validity comes from photographic thickness versus distance of plumes from three power plants in Australia (Carras and Williams, 1983). Good agreement with σ_z/z_i in Lamb's numerical experiments is found for the less buoyant plumes ($F_b < 300\ m^4/s^3$). Further testing of Willis and Deardorff's convective scaling results was recently accomplished in a wind tunnel, at Colorado State University (Poreh and Cermak, 1984). This was described in Section 3, under "Laboratory Experiments." Rather good agreement with the tank results was obtained, in spite of lack of complete development of convective turbulence intensity and other restrictions that accompany scaling-down to wind tunnel size.

At present most theoreticians seem convinced that convective scaling is the best way to go for modeling daytime, non-neutral diffusion ("the best thing since sliced bread" was heard in this context at a recent conference). Many modelers have been fitting the experimental results with analytical formulas for σ_y, σ_z, or surface $\int \chi dy$, using convective scaling. Others have developed fairly simple "ballistic" (or "bouncing ball") models for vertical diffusion in the CBL using skewed vertical velocity probability distributions. These attempts to develop applied models using the new information based on convective scaling were extensively reviewed at the U.S. EPA-AMS Conference on Updating Applied Diffusion Models (Briggs, 1985b).

Weil and Brower (1982) have suggested $w^*/\overline{u}$ as an index of diffusion rates for convective conditions. Since the Minnesota experiments and others have shown that $\sigma_v \simeq 0.6w^*$ throughout the mixing layer, and that σ_w attains nearly the same value through most of this layer (0.1 to 0.8 z_i), the initial spreading rate of a plume is approximately 0.6 $(w^*/\overline{u})x$ for either σ_z of an elevated source or σ_y for any source within the layer. This follows from statistical theory, since near the source $\sigma_y \simeq \sigma_a\ x \simeq (\overline{\sigma_v}/u)x$. Also, the descent of maximum concentrations from an elevated, passive source is proportional to $(w^*/\overline{u})x$. Since $\sigma_a \simeq 0.6\ w^*/\overline{u}$ in these conditions, this index is closely related to the Brookhaven classification system, but Weil and Brower use it more quantitatively, as a continuous index.

LARGE EDDY SIMULATION MODELS

Another numerical modeling approach to boundary layer turbulence that is really quite different from second order closure models has been called the "subgrid closure" or large eddy simulation (LES) model. This is the approach first applied long ago to prediction of long-range global weather patterns (Smagorinski, 1963), then further developed for the CBL by Deardorff (1970). The equations for the motions of fluids are believed to be complete and can be solved exactly for many non-turbulent flows. The "closure problem" arises with turbulent flows because the equations can be solved exactly only if initial and boundary conditions are known exactly and we attempt to calculate every detail of the motion, down to the smallest eddy. This, of course, exceeds the computational capacity of any computer so far envisioned (unless we regard wind tunnels and water channels as analog computers!). For tractability, we have to introduce some kind of averaging, and this always introduces some new unknown, such as entrainment velocity or shear stress. Some assumption must be made about how the new unknown(s) relate to quantities already computed to "close" the problem and make a solution possible. In the LES modeling approach the domain of computation is divided up into grid cells and the exact, or "primitive", equations are used to solve for the mean velocity, temperature, water vapor content, etc., of each grid cell for each time step. Thus, the details of the large eddy turbulent motions are actually computed in time and space; this gives a "picture" of the turbulence, if desired. Patterns of mean quantities are obtained by applying time averaging, just as would be done in a field or laboratory experiment. This contrasts to higher order closure models, which predict mean patterns but none of the details of motion. What is lacking, however, is exact computation of the way neighboring grid cells interact with each other through turbulent exchange of momentum, heat, water vapor, etc., by subgrid scale turbulent eddies. The closure problem is still there, but in this approach it has been pushed down to the scale of the grid cells used; this follows because bulk averaging is done on this scale with nothing known about the details of motion within the grid cells.

One advantage of this approach is that the structure of fine scale turbulence is simpler than that of the large scales where turbulent energy is produced. The small eddies are more isotropic and are driven only by the "inertial" forces induced by the larger eddies. This is called the "inertial subrange" of turbulence, and turbulent velocity spectra follow a simple -5/3 power law with frequency or wave number in this regime. This ought to make the closure problem simpler on this scale, with greater universality of successful solutions. The closure that has been used is a rather primitive eddy diffusivity assumption; the rate of subgrid turbulent exchange between cells is assumed proportional to the difference between the cells of the quantity under consideration. This seems a "safe" use of K theory because if the grid cells are small enough, the gradients of mean quantities do not vary much on that scale. K theory is used only to account for exchange done by eddies of the grid cell size and smaller.

The eddy diffusivity for subgrid exchange used by Smagorinski (1963), Lilly (1967), and Deardorff (1972) is based on the cell size and velocity differences between adjacent cells. In effect, it is assumed that the subgrid turbulence reacts instantly to changes in local velocity gradients. More specifically, the subgrid K is taken to be proportional to the square of the grid size times the local deformation rate, which is always positive. The constant of proportionality is chosen to produce flux-gradient relationships in agreement with those derived from careful field experiments, like the 1968 Kansas program. Deardorff (1972) uses an eddy diffusivity for heat, K_h, three times as large as the one for momentum, K_m (except for the layer of grid cells adjacent to the surface); K_h and K_m are different because of the pressure-velocity correlation. Pressure forces affect momentum exchange but not heat exchange. Presumedly, this applies to material exchange also, so on a subgrid scale the turbulent exchange of concentrations between cells should be described by K_h.

Deardorff (1972, 1974) first applied this type of model to ABLs for neutral and convective cases. Many of these results were mentioned in the previous subsection on convective scaling models, since they led to his convective scaling suggestion. One very instructive product of this model, not available with other models, is the mapping of fields of turbulent quantities computed for any given grid plane and time step. You can see the eddies, so to speak. It must be remembered that these are only model predictions, however, and it must be asked how well the model simulates what is known about the ABLs. Deardorff's model results have compared very well with aircraft and tethered balloon measurements for convective cases. The greatest shortfall has been in the magnitude of computed lateral turbulent velocities. This may be partly due to the limited width of the domain of computation, 2 z_i (the edges of the domain are not treated like solid walls; cyclic boundary conditions are used, so whatever advects in or out of one wall is balanced by advection through the opposite wall). The model is not capable of computing subgrid details like entrainment at the top of the mixing layer; instead, a "solid" lid at $z = z_i$ was used, with $w' = 0$ there.

Deardorff's model output for velocity fields was used directly by Lamb (1978, 1979) for numerical diffusion experiments. In effect, Lamb released thousands of numerical "particles" into the already-computed turbulent velocity fields, computed their trajectories, and counted them over grid volumes to get average concentration fields. He also added to the particle velocities a random subgrid scale velocity component, scaled by subgrid-scale turbulent energy. The results for release heights near the surface, at $z \simeq z_i/4$, and at $z \simeq z_i/2$ in convective cases are in satisfactory agreement with Willis and Deardorff's modeling tank results, as was discussed in the previous convective scaling subsection. They agree especially on the "unorthodox" qualitative behaviors of the surface and elevated plumes.

Large eddy models are at their best in large eddy turbulence, which is why the convective case has been the most explored. Deardorff also modeled the NBL. In principal, this type of model can also be applied to SBLs. A much larger computational capacity will be needed, however, because the largest eddies (for vertical turbulence, at least) are much smaller. The nocturnal mixing depth typically goes to 10 L, and eddy size is limited by L, rather than z_i. Instead of a computational domain 20 cells high, as Deardorff used, at least 100 are needed. Much smaller time steps would be needed also because the rotation period for the most energetic eddies scales to L/u^* instead of z_i/w^*, about an order of magnitude shorter time period. If Brost and Wyngaard (1978) are correct, the model would have to be run a much longer time because even on slight terrain slopes a steady state is not achieved during a night. In contrast, the CBL approaches steady state in its turbulent properties after a few z_i/w^*, typically 0.5 to 1 hr.

STATISTICAL MODELS

It was realized long ago that if enough were known about the statistics of turbulence and the mean flow field, it should be possible to calculate the dispersion of passive substances resulting from the turbulence. The formal foundation for this calculation was laid by G. I. Taylor in 1921 and is summarized by Pasquill (1974, see section 3.4). Its main limitation is that it applies only to homogeneous turbulence, i.e., the dispersant is assumed not to spread into a region where the statistics of turbulence are different from the origin. Thus, it applies best to horizontal diffusion over terrain flat and homogeneous enough to not cause large variations in the lateral turbulent velocity statistics. It can be applied to vertical diffusion close to an elevated source, but it becomes mathematically unsound as the material spreads and approaches the ground, since vertical turbulence statistics vary strongly with height near the ground. This is not to say that this application should not be tried; with appropriate modifications, useful results might be obtained, as is proven by data comparisons with the "bouncing ball" models mentioned in the subsection on convective scaling models (these models are based on turbulence statistics at just the source height). The statistical approach of Taylor has been extended and adapted for the case of vertical diffusion from a ground source (Hunt and Weber, 1979) by separating the motion into mean rise velocity, $d\overline{z}/dt$, and fluctuations about this mean and using the fact that $\sigma_w{}^2$ does not vary much with height.

The statistical theory of Taylor and the simplified approach to it developed by Hay and Pasquill (1959) need a capsule review here because they are so fundamental to all subsequent work with statistical models. Taylor's theory, as applied to turbulent diffusion, begins by describing the displacement of a single particle in terms of its time history of velocity, assuming that it follows the motion of a single parcel of fluid. This is the Lagrangian velocity, and all statistics of such velocities are called Lagrangian. A second particle can be added and the relative displacement can be described in terms of the Lagrangian statistics of the two particles to compute the relative diffusion. The most frequent application of this theory is to total diffusion, i.e., the variance of displacement of a large number of particles released steadily from the same point. One fundamental result, written here for only the lateral component of diffusion, is

$$d\sigma_y^2/dt = 2 \int_0^t \overline{v'(t)v''(t + \xi)\, d\xi} \qquad (5.5)$$

$$= 2\, \sigma_v^2 \int_0^t R(\xi)\, d\xi \quad ,$$

where t is the travel time, σ_v is the total velocity variance, and $R(\xi)$ is the Lagrangian correlation coefficient of velocity separated by a time lag ξ. Since $R(\xi) = 1$ at small ξ, i.e., there is no significant change in parcel velocity over a small time, it follows from the above that $\sigma_y = \sigma_v t$. For very large travel times, if $R(\xi)$ approaches zero faster than ξ^{-1} at large ξ, $\sigma_y = (2\sigma_v^2 t_L)^{1/2}$, where t_L is the limit of the integral of $R(\xi)$ over a large time; t_L is the "Lagrangian time scale". This is the same result given by constant eddy diffusivity, with $K = \sigma_v^2 t_L$. Correlation coefficients have a direct mathematical relationship with spectra, so the basic theory of Taylor was later developed in terms of integrals of Lagrangian velocity spectra by Batchelor (1949) and others.

These results are exact and relatively simple, but true Lagrangian velocity statistics are rarely obtainable. In an approximate way, large scale horizontal velocities can be deduced from the trajectories of tracked balloons, but these never follow the vertical air motions perfectly. What is needed are cheap, small, neutrally buoyant air motion "trackers". Because it is much easier to obtain Eulerian velocity statistics, which are measured at a fixed point such as a meteorological tower, Hay and Pasquill (1959) tried a different approach. They assumed that the Lagrangian velocity correlation, $R_L(\xi)$, is similar in shape to the

Eulerian velocity correlation, $R_E(t)$, except for an expansion of the time scale by a factor β; that is, $R_L(\xi) = R_E(\xi/\beta)$. Pasquill (1974) has argued that the exact shape of the velocity correlation is not critical anyway. The value chosen for β is more important. One simple estimate for β is based on a scaling approximation for the two time scales. The Eulerian time scale, t_E, is proportional to the time it takes an eddy to pass a fixed point, which is $\ell/\overline{u}$, where ℓ is the size of the dominant eddies. The Lagrangian time scale is proportional to the time it takes to traverse an eddy with a turbulent motion, ℓ/σ_v, if lateral diffusion is being considered. Then $\beta \propto t_L/t_E \propto \overline{u}/\sigma_v = i^{-1}$, where i is called the turbulence intensity. Pasquill (1974) gives estimates of (βi) ranging from 0.35 to 0.8.

The most useful result so far from the Hay-Pasquill hypothesis has been the following relationship, written here for only the lateral component of spread,

$$(\sigma_y)_\tau = (\sigma_v)_{\tau,t/\beta} t \simeq (\sigma_a)_{\tau,t/\beta} x. \tag{5.6}$$

Here, we use the approximations $\sigma_a \simeq \sigma_v/\overline{u}$ for the azimuth angle variance and $t \simeq x/\overline{u}$ for the time of travel; τ is the sampling duration; and t/β is the averaging time, which acts like a filter. By smoothing over an averaging time t/β before calculating variances, the contribution to σ_v of high frequency eddies that only move the particles back and forth a number of times by the time they reach x is suppressed. Using the above, all that is needed to calculate dispersion in homogeneous turbulence is a record of the velocity or azimuth angle fluctuations at the release point plus a good estimate of β; this method can be applied to different sampling durations in a straightforward manner. This approach is, in fact, the main basis for Pasquill's curves for σ_y, rather than direct diffusion measurements. Pasquill (1974) has accumulated much experimental support for this approach, especially for σ_y.

Statistical theory goes back over 60 years, yet, since 1976, there has been a marked increase of interest in it as a practical tool for diffusion modeling. The first of these recent attempts to make practical simplifications in statistical theory was by Draxler (1976), who used the forms suggested by Pasquill (1971)

$$\sigma_y = \sigma_v t \; f_1(t/t_L) \quad \text{and} \tag{5.7}$$

$$\sigma_z = \sigma_w t \; f_2(t/t_L) \quad , \tag{5.8}$$

where σ_v or σ_w are measured over the same sampling duration as σ_y or σ_z, with an averaging time approaching zero (i.e., unfiltered fast-response measurements). Draxler attempted empirical evaluations of f_1, f_2, and t_L separately for elevated and surface sources, horizontal and vertical diffusion, and stable and unstable conditions. Altogether, 11 different sets of field experiment observations were used. Analytical best fits for f_1 and f_2 were suggested as were best values of $T_{1/2}$ for each category, where $T_{1/2}$ is the time that f_1 or f_2 is observed to drop to 1/2 (the derived value of t_L was 1.64 $T_{1/2}$ for most cases). $T_{1/2}$ ranged from 50 s for σ_z of ground sources in stable conditions to 1000 s for σ_y of elevated sources for all conditions. However, the scatter in individual run best fits to $T_{1/2}$ was very large, over several orders of magnitude. The scatter in the logarithms of the ratio of computed to observed σ_y and σ_z, using the suggested $T_{1/2}$ values, ranged from a standard deviation of 1.29 to 2.05. These were somewhat better than the comparison with the Pasquill-Gifford predictions, but still represent errors of factors of 4 to 8 - not a very comfortable place to rest the "state-of-the-art". Nevertheless, in 1977 this approach was recommended by an AMS workshop as a step toward improved modeling (Hanna et al., 1977). For σ_y, the simple form suggested by Pasquill (1976a) was also recommended, with $\sigma_y/(\sigma_a x)$ a function of only x. It should be recognized that much of this scatter is due to poorer instrumentation for measuring σ_a and σ_e in the older experiments. Draxler's method performed much better when compared to recent experiments (Gryning and Lyck, 1984).

Subsequently, there have been attempts to reduce the amount of pure empiricism in this approach, to better model the variability of quantities like $\sigma_y/(\sigma_a x)$. Doran et al. (1978) showed how this quantity, plotted against x, systematically depends on the sampling duration and the averaging time used in different experiments. They qualitatively related this dependence to velocity spectra considerations and suggested that wind speed and stability will also have important effects. The same authors studied additional data bases and concluded that σ_a measured at elevated release points is not a good indicator of surface level σ_y; near-surface σ_a orders the diffusion data far better (Horst et al., 1979). Examining data for rough terrain experiments, they found indications of problems with σ_a measurements; in the older experiments, wind vanes may have been either sluggish or insufficiently damped.

Irwin (1979) suggested some forms for $\sigma_z/(\sigma_e x)$ in which σ_e is the variance of the elevation angle. For neutral and stable conditions he presented them as empirical functions of travel time, with one curve for $z_s > 100$ m and another for $z_s < 50$ m, and a suggested interpolation; these were based on averages from five experiments. For unstable conditions, he made the physically reasonable assumption that t_L is proportional to z_i/w^* and presented the Willis and Deardorff tank data, Lamb's numerical results, and some limited field data as functions of tw^*/z_i, the convective scaling dimensionless time scale. Separate curves were fit to ground source and elevated source data. This approach does appear to give considerably less scatter than the purely empirical analyses.

A somewhat similar scientific approach was recently used for elevated source values measured in a SBL (Venkatram et al., 1983). The data were lidar measurements of σ_z from an oil fog plume released from a crane (Lavery et al., 1982). Instead of $\sigma_e x$, $(\sigma_w/\overline{u})x$ was used to nondimensionalize σ_z for plotting against t/t_L. For t_L, an argument was made that in the SBL $t_L \propto N^{-1}$, where N is the Brunt-Vaisala frequency of oscillation defined by the potential temperature gradient $N^2 = (g/\overline{\theta})\ \partial\theta/\partial z$. The results were good for this data set; a good fit was found to the form for σ_z suggested by Draxler (1976). In a subsequent development (Venkatram et al., 1984), an improved fit to observations from the same experiment was found using $t_L^{-1} = 3.7N + 2.8\ \sigma_w/z_s$; all meteorological measurements were referred to the source height, z_s. This formulation includes the neutral limit, $t_L \propto z_s/\sigma_w \propto z_s/u^*$. An impressive 95% of the observations fell within a factor of 2 of the curve $\sigma_z = \sigma_w t(1 + t/2t_L)^{-1/2}$. However, in the upper half-decade of their t/t_L values, some tendency towards a -1 slope in the plot for $\sigma_z/(\sigma_w t)$ can be seen. This suggests $\sigma_z \propto \sigma_w t_L$ as a possible limit, a limit that has been suggested by Hunt (1982) for the case $t_L \propto N^{-1}$.

The effect of sampling duration on plume width and average concentration can be readily determined by using statistical theory, a capability that escapes most other modeling approaches. All that is needed is a set of velocity spectra or a time series of fast-response wind speed or directions from which quantities like $(\sigma_a)_{\tau,t/\beta}$ can be computed. A classical study was done by Hino (1968), who considered several spectral forms as well as wind records analyzed using the Hay-Pasquill assumption. With some differences in predicted or observed ranges of application, he found that a maximum concentration $\propto \tau^{-1/2}$ was a practical approximation to most model estimates and observations. The dependence on τ was weaker for $\tau < 10$ min, especially with small travel time, t. Hino noted from his analysis of wind observations in Japan that concentrations drop off more strongly than $\tau^{-1/2}$ for τ greater than about 3 hr in neutral conditions, but that $\tau^{-1/2}$ applies out to 5 hr or more in unstable conditions. The statistical theory could be applied in this way to considerably more detailed studies of the effects of sampling duration on maximum concentrations; there could be separate analyses for σ_y and σ_z, various heights, and several ranges of stability. Travel time or distance is also a factor. All that is needed is a meteorological tower with recording, fast-response wind sensors; even better would be use of several such towers in different terrain types.

"Cluster growth", or relative diffusion, is another problem area that, in principle, is amenable to solutions using statistical theory. In practice, this is a much more difficult problem to solve because one must know or assume properties of relative Lagrangian velocity statistics, which depend on the spacial separation of particles as well as time lags. Taylor first addressed the problem (Batchelor, 1952), and Smith and Hay (1961) attempted to simplify it by replacing Lagrangian velocity correlations with Eulerian ones, in the manner of Hay and Pasquill (1959). This theoretical matter then seemed to get little attention until Sawford (1982a) discussed and compared both approaches. Data for testing predictions for relative diffusion were limited and contained much scatter, but Sawford believed that Taylor's approximation gives more realistic predictions.

RANDOM PERTURBATION MODELS

In the last few years there has been a resurgence of interest among theoreticians in mathematical simulations of turbulence diffusion that use randomness as one element of the simulation. These are variously called random walk, Markovian random walk, Monte Carlo, random force, and Langevin equation methods. Much of the mathematics of these methods has been borrowed from treatments of Brownian motion, which were made much earlier in the century by a number of prominent scientists, e.g., Einstein (1905).

Most of these methods also lean heavily on statistical approaches to diffusion modeling, since they depend on velocity correlation functions for the non-random part of the modeling. An exception is Lamb's (1978) modeling of the CBL, which uses Deardorff's (1973) large-eddy, subgrid-closure model to determine motions caused by large eddies and randomized velocities for diffusion caused by eddies on the subgrid scale. Undoubtedly, computational advances account for some of the renewed interest in these methods, especially the Monte Carlo techniques that "compute" the trajectories of thousands of particles using random number generators.

The most common starting point is the assumption that

$$v(t + \Delta t) = R(\Delta t)\, v(t) + v'(t) \qquad (5.9)$$

(Smith, 1968), where v is the velocity of a small volume of air, Δt is a small increment of time, and v' is a random change in the velocity, with $\overline{v'} = 0$ and $\overline{v'v} = 0$. It follows that for stationary, homogeneous flow,

$R(\Delta t)$ must be the Lagrangian autocorrelation function (Gifford, 1982). Another form of the above uses the series expansion of $v(t + \Delta t)$ in terms of Δt, making Δt small enough that higher order terms can be neglected, so that $v(t + \Delta t) = v(t) + \Delta t dv/dt$. Substitution into the above yields

$$dv/dt + [1 - R(\Delta t)]\Delta t^{-1}v(t) = v'/\Delta t. \qquad (5.10)$$

This is a form of Langevin's equation, which was widely utilized in theories of Brownian motions; some of these theories are referenced by Gifford (1982), who also mentions several applications of the equation to turbulent diffusion made in the early 1960s.

The above equations, while simple, have many interesting properties. If the process is stationary, $\overline{v^2}$ must be constant in time; this imposes the condition that $\overline{v'^2} = \overline{v^2}[1 - (R(\Delta t))^2]$, so that the random velocity changes do not cause an increase or decrease in the total averaged kinetic energy. Quite a few investigators have assumed that $R(\Delta t)$ has an exponential form, as a matter of convenience; Gifford (1982, 1983) claims that this form of R can be derived from the fundamental equation. On the other hand, Pasquill and Smith (1983) claim consistency of this equation with any form of $R(\Delta t)$ that decreases linearly in Δt for small Δt; they also fit nonlinear forms of R, $R = 1 - \partial\Delta t^n$ for small Δt with $n \neq 1$, to more complex velocity change assumptions involving v at time steps prior to t (double regression and triple regression models). However, the exponential form is the simplest, and it has been claimed that diffusion predictions are not sensitive to the exact form of R (e.g., Pasquill, 1974). Smith (1968) notes that the exponential form for R implies impulsive forces acting on the parcel, as the analogy to Brownian motion would suggest, but this may be of little consequence to the practical outcome, the diffusion prediction.

Some controversy exists over the valid applications of this assumption to diffusion prediction. Gifford (1982) applied the Langevin equation, with $R = \exp-(\beta t)$, to derive analytical predictions of the mean displacement of particles with an initial velocity v_0 and the mean-square displacement of the same particles, σ_y^2. Applied over all initial values of v_0, with $\overline{v_0^2} = \overline{v^2} = \sigma_v^2$, he derived the same equation for total variance that Taylor (1921) derived from statistical theory (Taylor also assumed the exponential form of R). Applied to small initial values of v_0, Gifford showed that the results approximate $\sigma_y^2 = v_0^2 t^2$ at small t, $(2/3)\sigma_v^2 \beta t^3$ at intermediate t, and $2\ \sigma_v^2 \beta^{-1}$ at large t (β^{-1} is the Lagrangian time scale of the turbulence and $\sigma_v^2 \beta^{-1}$ is the same as eddy diffusivity, K). He pointed out that this t^2-t^3-t behavior of σ_y^2 contained all the asymptotes of classical statistical theory for both relative ("puff") diffusion and time-averaged diffusion. With appropriate choices of K, β^{-1}, and v_0, he was able to fit the prediction to a wide range of σ_y data, over travel times ranging from a few seconds to a year (radioisotopes from a nuclear test). Smith (1983) disputed the part of Gifford's interpretation having to do with puff growth, or relative diffusion. The debate is not easy to follow, but Gifford's side of it seems to be that since an instantaneous puff would have some initial velocity, v_0, whatever happened at the time of release, the σ_y prediction could be appropriate for puff diffusion if a typical v_0 is chosen. Smith's interpretation is that the σ_y equation applies to the averaged diffusion of all puffs leaving the source with the velocity v_0, not to the average relative of an isolated puff diffusion.

Although we must be careful about the interpretation of various results, the Langevin equation with an exponential form of R is very mathematically malleable. It was applied to relative and total diffusion from a finite-sized, finite-duration source by Lee and Stone (1983a). They also computed peak to mean concentration for a point source. A comparison paper explores relationships between Eulerian and Lagrangian time scales, so important in statistical diffusion theory, using the Langevin equation with Monte Carlo simulations (Lee and Stone, 1983a). Sawford (1982) used a similar method, except that he applied it to the relative motion between a pair of particles, a more fundamental approach to the problem of relative diffusion.

Most of the above results apply to horizontal diffusion because of the assumption of homogeneous turbulence; this assumption is seldom justifiable in the vertical dimension. Ways to extend the application of the method to the case of a gradient in turbulent kinetic energy were developed by Wilson et al. (1982) and Ley and Thomson (1983); this requires compensation for a bias velocity that develops in the direction of the gradient. Some conclusions from these studies are given by Pasquill and Smith (1983); one is that about 5000 random-walk trajectories must be calculated to get smooth profiles. Davis (1983) used these simulation methods to obtain a number of practical results for elevated-source diffusion in neutral conditions. He explored the effects of varying the release height, the surface roughness, and the friction velocity by making Monte Carlo experiments for a number of parameter combinations.

Very recent work on random walk modeling in inhomogeneous flows (Thomson, 1984) and inhomogeneous and unsteady flows (van Dop, Nieuwstadt, and Hunt, 1985) has shown that the Langevin equation can be used under these conditions, but a careful analysis is necessary near boundaries.

A more elaborate Monte Carlo scheme using Eq. 5.9 was developed by Baerentsen and Berkowicz (1984) for vertical diffusion in convective conditions. Recognizing the highly skewed nature of the vertical velocity distribution owing to the larger fraction of horizontal area occupied by downdrafts, they assumed separate statistics for updrafts and downdrafts, in effect, a double Gaussian distribution of w. Particles in an updraft or downdraft may stay in it until reaching $z = z_i$ or $z = 0$; then, perfect reflection is assumed, with a switch in the updraft/downdraft category that selects the statistical parameters. A particle traveling in an updraft, for instance, rises with the mean velocity of updrafts at that height plus a deviation velocity based on the Monto Carlo method and the mean variance of updraft velocities at the height being traversed. All velocity statistics were estimated, in terms of convective scaling, on the basis of field data, water tank data (Willis and Deardorff, 1974), and Lamb's (1982) numerical modeling. The Monte Carlo scheme provided a very good duplication of the vertical diffusion results from the tank experiments (Willis and Deardorff, 1976, 1978, 1981).

The computation of thousands of particle trajectories of the Monte Carlo method can be avoided by means of the "integral equation" approach, which integrates all probabilities mathematically by considering the probability of a "collision" (change in v) in each grid square (Smith and Thomson, 1984). This requires much less computation than the Monte Carlo method. With either approach, it is relatively easy for today's computers to use the random perturbation methods to calculate diffusion; once confidence in the assumed turbulence parameterizations is established, these methods can be used to explore the effects of release height, roughness, stability, etc, much more easily than by doing field experiments for every combination of circumstances.

SECTION 6

WHAT ARE THE NEEDS?

Perceived data and modeling deficiencies

EXPERIMENTAL VERSUS MODELING NEEDS

If there is a overwhelming scientific consensus on anything, it is that we need well-conceived new experiments much more than we need more models. This is not to say that there are no needs for new models or model development. There have been frequent calls to replace the P/G/T curves, especially for elevated-source modeling, and to replace the Gaussian plume assumption in situations in which it seems dubious, especially for vertical diffusion in convective conditions. However, alternative models are available, as Section 5 of this paper amply shows. The real need is to prove that, at least, some of them work better than models in use, and this requires a broader and more reliable experimental base than we have at present. Comparative studies like Irwin's (1983) are handicapped by the doubtful quality of wind variance measurements in pre-1970s experiments, the likelihood of significant deposition losses in the same, the sparsity of direct vertical diffusion measurements, and the rapid decline in data for distances greater than 1 or 2 km downwind. In addition, scientists have often expressed alarm and amazement at how far accepted modeling schemes extrapolate beyond the data bases supporting them; Smith (1981) compares this practice with extrapolating the parameters for a two-story house to construct a building the size of the World Trade Center. This is hardly an exaggeration; P/G/T σ_z values based on Prairie Grass measurements to 800 m have been extrapolated to 100,000 m. More extended experiments are needed both to support better models and to evaluate the adequacy of models that have been used for over 20 years.

A glance at the reference list for Table 1 supports the claim of Gifford (p. 4 of Smith, 1981) that experimental efforts dried up after 1968, at least for field experiments (the Hanford-67 series is an exception, spanning 1967 to 1973). He and Smith (1981) attributed this to a misplaced faith in numbers produced by computers, a commodity that became ever more abundant and cheap. In the period following 1968, considerably better tracers, sampling equipment, remote sensing techniques, and turbulence instrumentation were developed. So we are now in a position (if we had the money) to do much higher quality diffusion experiments. The recent elevated-source experiments in Europe, in Copenhagen and Cabauw, give us some encouragement, as does the CONDORS experiment in Colorado (Agterberg et al., 1983; Gryning, 1981; Moninger et al., 1983). These were all limited objective, but high quality, experiments. There has been moderate, continuing support for laboratory diffusion experiments through the 1970s up to the present. This support, however, has shifted toward studies of complex terrain, acidic deposition and buoyancy effects; there is little work at present on basic diffusion phenomena.

On the model development side, the opinion expressed by Smith (1981) that it needs no new support seems extreme; nor do we share his confidence that "new sources of field data will be devoured by the modelers as quickly as they become available." Although computer power becomes steadily cheaper, it takes considerable human effort to assemble all the new and old pieces of information, process it effectively using computers, and intelligently evaluate the results, including needs for model revisions. Comparison of one model with one experiment is a relatively modest task, but much more is learned, and more effort is required, when many models are compared with that experiment (e.g., Gryning and Lyck, 1984). Best yet are efforts to compare many models with many experiments (Turner and Irwin, 1982; Irwin, 1984). These efforts require at least the current level of support for model evaluation, as most scientists engaged in this activity feel that it will take a long time to meet current needs at the current rate of commitment. Furthermore, it takes much well thought out effort to translate research models, once validated, into appliable models, given the fact that we do not normally have high-grade meteorological measurements available. This need will be addressed at length in Section 7. The need for better applied models is indisputable. The scientists participating in the Rural Model Review were critical of the eight models reviewed as being too similar and showing too little skill (Fox et al., 1983). The AMS committee conducting this review strongly urged that "the scientific community submit models that it considers technically better than those available today."

Because modeling improvements require suitable data bases for validation, specific modeling needs and experimental needs tend to parallel each other. Specific areas needing attention will be addressed in the following subsections.

WHAT KIND OF EXPERIMENTS ARE NEEDED?

Suggestions for the kind of experiments that are needed for improvements in diffusion modeling are numerous. This section will discuss generalities, leaving specific phenomena needing investigation to the sections on the SBL, the CBL, and on "calms and other inconvenient events".

One thing is clear: scientists want diffusion experiments with a complete complement of measurements. It is frustrating to be unable to use diffusion measurements to test a model because some critical meteorological variable was not measured. Sometimes a climatological or other estimate of the missing variable can be substituted, but this muddies the analysis, making the results less certain. There are many past experiments that give only σ_a and/or σ_e; there is no hint of even the P/G/T stability. In either case, no matter how high the quality of the diffusion measurements, the data are useful only for testing a certain subset of diffusion models. Considering the trivial added cost and bother of adding "cloud cover" to the list of measured variables, this is ridiculous. A number of other meteorological parameters are more difficult to obtain, but nevertheless are quite cost-effective in maximizing the usefulness of the experiment. The goal should be to produce a set of measurements that can be used to test any physically reasonable diffusion model, present or future. The Prairie Grass experiment is an outstanding, solitary example of what can be done; every possibly relevant meteorological variable that could be measured in 1956 was measured. Who envisioned then that the aircraft soundings to 3000 m would be useful for determining

z_i and w^*, when convective scaling theory was developed in 1970? This vintage experiment has been more useful than any other, even in model evaluations of the most recent decade.

There is also a strong consensus that diffusion measurements are needed at much larger x than in past experiments, e.g., see Ching et al. (1983); no one is comfortable with extrapolations to 100 km based on data to about 3 km. Many scientists mention 100 km as a goal distance, as recommended by the AMS Review Panel on Sigma Computations (Randerson, 1979). Especially needed are direct measurements of σ_z or of vertical concentration profiles; in the CBLs, this need extends only to distances where the material becomes well-mixed from $z = 0$ to z_i, say to $x \simeq 4\,\overline{u}\,z_i/w^*$. In neutral and stable conditions, there is little idea of the distances at which vertical growth terminates. There is also uncertainty about σ_y at large x; does it approach an asymptote $\propto x^{1/2}$, $\propto x^1$, or something in between? Extrapolating from 3 to 100 km, the choice of asymptote can make quite a difference. There are many who believe that wind direction shear is an important factor in σ_y growth beyond $x = 10$ km; thus, $\theta_a(z)$ should be measured through the depth of the boundary layer containing the plume.

There have also been a number of calls for experiments over some variety of terrain (e.g., Randerson, 1979). The Prairie Grass experiment, while very valuable, represented an extremely flat, smooth terrain. The terrains in which most people live and most pollution problems occur lie broadly between the prairie field of wheat stubble and "complex" terrain. There is a need for experiments in areas with typical surface roughness, with trees, buildings, and scattered clearings (the Copenhagen and Cabauw experiments, Table 1, are good examples). There is also a need for experiments in rolling or moderately hilly areas, especially in stable conditions, when drainage flow effects develop even on the slightest of slopes.

The need for plume measurements at larger distances parallels a need for more complete plume characterization. A sparse sampler network merely results in questions about representativeness, whether the peak concentration was measured, and where the centerline of the plume was located. Where it is permissible to create a highly visible plume, even old-fashioned smoke photography can produce more useful results, giving both instantaneous and time-averaged plume growth (Nappo, 1984). This is not feasible for large x, however, except in stable conditions at a site remote from development. Remote sensors, especially lidar, are seen by many as the most satisfactory way to get three-dimensional plume measurements. Lidar can either be surface-based or mounted on a plane traversing a plume from overhead (Johnson, 1982). Lidar signal processing techniques have now developed to the point that χ/Q isopleths can be mapped (Eberhard et al., 1985). The overhead "burden" ($\int \chi dz$) of SO_2 and other gases can be measured by van mounted correlation spectrometers, and the ground level concentrations can be measured with fast-response gas analyzers by using four to six traverses per hour across a plume on a highway; this technique has been used to define power plant plumes at distances up to 25 km (Weil, 1977). For much larger distances, several excellent tracers have been developed that do not significantly deposit and that can be detected at great distances, up to x = 1000 km (Johnson, 1982). The logistics of operating and maintaining a sampler network on such scales is difficult and considerably increases the cost of an experiment.

A few other experimental needs have been mentioned. One is for more elevated-release experiments, with larger z_s, typical of the taller sources we are modeling. The recent experiments discussed in Section 3 had source heights ranging from 115 to 285 m, so there has been some progress for more elevated release experiments. The majority of releases from taller sources are also quite buoyant, which involves a new range of phenomena that is best studied with buoyant plumes as the objects of measurements, rather than passive releases. One very good idea, at least if kept on a scale that makes it not too expensive, is a multiple day around-the-clock experiment such as done in the Tennessee Plume Study (Schiermeier et al., 1979). Past experimenters have sought short periods of steady-state meteorology in order to get data in the ideal conditions assumed by modelers. The atmosphere is never in a steady state for long, and many modelers and users wonder if transition periods, turbulent "bursts" in the SBL, and other transitory phenomena are being mismodeled.

The above discussion has concerned needs for field studies. While some phenomena, like wind direction shear effects and large-x diffusion, may not be practical objects of laboratory modeling studies, wind tunnels, water channels, and tanks are very useful for studying short-range phenomena. Since they are cheaper to operate than field studies, they ought to be more fully exploited. A few suggestions regarding these studies were made at the end of Section 3.

METEOROLOGICAL MEASUREMENTS

As mentioned in the previous subsection, diffusion field experiments cannot be fully utilized for testing models unless they are accompanied by complete meteorological measurements. There are also practical needs concerning meteorological measurements to be used in applied modeling. This subsection addresses both of these concerns.

The meteorological parameters needed for testing current research models are σ_a and σ_e at source height, u^*, L, and z_0 and, for daytime, z_i and w^* (there is one redundancy among these, because $w^{*3}/z_i = u^{*3}/kL$). For large-x, lateral diffusion experiments it is important to measure mean wind direction, D, through the layer containing the plume, so that the rate of change of wind direction with height can be correlated with σ_y growth. In stable conditions, potential temperature profiles, $\theta(z)$, should be measured through the layer containing the plume because an important time scale affecting turbulence and vertical diffusion is N^{-1}, where $N^2 = (g/\overline{\theta})\partial\theta/\partial z$ (a well-established flux-profile law relating N^2 to u^* and L may seem to obviate this need, but the law applies only in the lower part of the SBL; in the middle to upper levels, there is less certainty about the relationship). For elevated releases in the SBL, it is very important to know the height of turbulent mixing, h, because vertical mixing is nil above it. Of course, wind speed is a primary parameter in all diffusion models, so it at least must be measured at some height representative of the plume; better, $\overline{u}(z)$ profiles should be measured through the layer containing the plume. It is also desirable to have more than a minimal number of $\overline{u}$, $\overline{\theta}_a$, and T measurement levels in case an instrument becomes impaired during the course of the experiment, which is a strong possibility.

These measurements can be made roughly or with high quality instrumentation, depending on the overall level of effort in the experimental program. The surface roughness, z_0, can be crudely estimated by eye, or it can be more properly determined from wind profiles taken near the ground in near-neutral conditions. An even better method is given by Wieringa (1976, 1980). This estimate of roughness uses the longitudinal turbulence intensity and gives an effective roughness. The surface flux quantities u* and L require fast response measurements of u, w, and T to be measured directly, and such measurements require some experience in quality assurance. The next step down is to estimate u* and L from established flux-profile relationships and profiles of $\overline{u}$ and θ; this approach has often been used in analyses of the Prairie Grass experiments, as excellent profile measurements were made on a 16-m tower, but fast-response turbulence instrumentation did not exist in 1956. The minimum measurements for estimating L and u* are the bulk Richardson number, $\overline{u}^2/\Delta\theta$, and z_0. Particular care should be taken in making these measurements; $\overline{u}$ should be measured near z = 10 m, and measurement between about 2 m and 10 m has been recommended for $\Delta\theta$ (Hoffnagle et al., 1981). Potential temperature is then calculated as a direct function of temperature and ambient pressure. The lateral and vertical wind direction variances can be replaced by the velocity variances σ_v and σ_w, respectively, by using $\sigma_a \simeq \sigma_v/\overline{u}$ and $\sigma_e \simeq \sigma_w/\overline{u}$. If u*, L, and z_i or h are measured, it is now possible to estimate σ_v and σ_w from these parameters on the basis of boundary layer research, with the exception of σ_v in stable conditions (large horizontal eddies may depend more on the terrain than on surface fluxes in stable conditions). However, the atmosphere does vary from its own norm, so it is much better to measure σ_v or σ_a and σ_w or σ_e directly.

For experiments in the daytime, z_i, must be measured by using lidar, rawinsondes, or other profile measurements that range to heights of at least 2 km in the summer (z_i frequently reaches 4 km over arid inland areas). For model applications z_i can be estimated from morning soundings at National Weather Service stations and insolation or surface temperature, but it is too important a parameter to estimate in such a manner for a research experiment. Likewise, for nocturnal experiments, h determines whether or not vertical turbulent mixing will occur at all for an elevated source and strongly affects the diminishment of turbulence velocities with height. Therefore it is an essential measurement. In the most common range from about 50 to 500 m, h can be detected with acoustic sounders (sodar); however, caution is necessary in interpreting the sodar signal. In very stable conditions, h can get as low as just a few meters deep. Turbulence measurements on a tower will indicate the heights above h by smooth traces on a strip chart. Cheaper means of detection would be movies of flag motion or small smoke releases at various tower heights or vertical traverses with one very responsive turbulence instrument.

In making the sophisticated measurements, one should not overlook the simple ones. To determine the P/G/T category, all that is needed besides $\overline{u}$ is a measurement of insolation, net radiation, or a cloudiness estimate. For slightly more refined systems, some notation of surface moisture (dry, lush plant cover, dew, soil set from recent rain, etc.) and a relative humidity measurement are needed.

This relates to another experimental need. Little work has been done on how to get the best estimates of difficult to measure parameters that are from relatively simple measurements. This must be done if research diffusion models are to be usable in improving operational models. Some work along this line will be discussed in Section 7; more work is needed. It is especially valuable to try out the simpler measurement schemes in the context of a diffusion experiment, so that we can discover how much degradation of diffusion model performance occurs when we use simple meteorological measurements as proxies for the difficult ones. Can a research model applied in such a way beat out the P/G/T or Brookhaven systems without a large increase in measurement effort? One step in this direction was made by Weil and Brower (1984) in the context of power plant plume modeling. They demonstrated that a Gaussian model using improved dispersion parameters outperforms a current operational model. Briggs (1982) showed that the Prairie Grass values of $u_*\int \chi dy/Q$ correlate about as well with simple substitutes for L as with L itself, the best measure in previous comparisons (Briggs and McDonald, 1978). The substitutes were $\overline{u}^3/R_s{}^*$ and $\overline{u}^3/R_n{}^*$, where $R_s{}^*$ and $R_n{}^*$ are proportional to solar insolation rate and net radiation, respectively, rather easily measured quantities. Even $\overline{u^3}$ alone provided useful correlation, but with increased scatter.

Many scientists have been thinking about the need to get the best estimates of quantities like u* and L for operational modeling purposes. Another concern is to get representativeness, because L can vary by a factor of 3 or more from field to forest (in this regard, mean quantities like $\overline{u^3}$ have more spatial averaging in them than quantities like the turbulent heat flux. The AMS Workshop on Stability Classification Schemes and Sigma Curves made recommendations on instrumentation for diffusion modeling (Hanna et al., 1977). More detailed recommendations resulted from the U.S. EPA Workshop on On-Site Meteorological Instrumentation Requirements to Characterize Diffusion from Point Sources (Strimaitis et al., 1981); this report included recommendations on instrument response specifications. Recent comparisons of turbulence velocity responsiveness and accuracy of six commercial systems were made by Kaimal et al. (1984a); comparisons were also made of four Doppler acoustic sounding systems capable of remote measurements of wind speed, direction, vertical velocity, and vertical velocity variance, σ_w (Kaimal et al., 1984b).

EVALUATION AND CONCENTRATION FLUCTUATION ISSUES

A surprising number of scientists are calling for more research on fluctuations of concentrations about their mean value in steady conditions, $\overline{\chi}$ (defined $\chi' = \chi - \overline{\chi}$). It is the nature of turbulence to be very variable even when the conditions producing it are steady (for example, note the gustiness a few feet in front of a steadily rotating electric fan). Given this natural variability, the question exists of how to know when $\overline{\chi}$, or $\overline{u}$, or $\overline{v'w'}$, etc, has been determined. The answer is that you cannot measure a "true" average quantity, but the longer you average in steady conditions the better the precision of the measurement. The problem in the atmosphere is that "steady" driving forces do not remain so for very long, so there is a limit to how precisely $\overline{\chi}$ and and mean quantities can be measured no matter how good the instrumentation (in a wind tunnel, you can maintain steady flow conditions as long as your budget allows, so you can get much closer to true averages). However, if one can measure the statistics of the fluctuating quantities, it is possible to state the probability of the resulting finite-time average being within a percentage of the true average. This should be a major consideration in the evaluation of the worth of experiments. (For instance, each Prairie Grass run consisted of only a 10-min release of tracer; because large convective eddies take several minutes to advect past a source, individual daytime trials showed large scatter with regard to one another, regardless of the parameterization scheme being tested. Scatter can be reduced by combining runs with similar parameter values.) At the other end of the modeling process, the consequence of turbulent variability is that even with the perfect model and perfect input parameters, a perfect prediction of the results that will occur for a particular hour cannot be made. Just as the weather is rarely "average" for the date, time, and place, atmospheric diffusion routinely deviates about its own mean, even given the same set of driving forces.

Thus, knowledge of concentration variability is essential to rational model evaluation. Given that even the perfect model cannot be in perfect agreement with data, except occasionally on chance, what degree of agreement should be our goal, and what degree of disagreement really indicates modeling deficiencies? This is the main motivation for recent interest in χ' measurements and models, as there has been much interest in how to properly evaluate diffusion models and how to specify error bounds in models (Fox, 1981; Fox et al., 1983). Other practical needs for χ' include prediction of flamability of accidental gas releases, the response of insect pests to smells locating their host plants, and prediction of lethal doses of fast-acting toxic gases. The U.S. EPA-AMS Workshop on Updating Applied Diffusion Models (Weil, 1985) focused on this issue. This workshop recommended pursuit of several χ' modeling approaches that appear promising, including second order closure models, large eddy models, and random perturbation (Monte Carlo) models (these are described in Section 5). The greatest needs are for χ' data. Laboratory simulations were recommended as valuable, as χ' measurements are easier to make in a wind tunnel or water tank than in the field, and $\overline{\chi}$ can be determined more accurately by extending the "steady" conditions. Recent work by Willis and Deardorff (1984) in their modeling tank began exploration on χ'/χ for buoyant plumes in a CBL.

A related need is to find the most effective ways to evaluate diffusion models with field data, which are always flawed and have unknown deviations from the true $\overline{\chi}$. The AMS committee overseeing the Rural Model Reviews was obviously quite disappointed in the results of extensive statistical comparisons between eight models and sampler data from one power plant; scatter of predicted-to-observed χ ratios was so large that none of the sophisticated statistical tests detected significant differences in model performance (Fox et al., 1983). Using a much simpler statistic, the fractional error, and stratifying the same data by distance and stability, Irwin and Smith (1984) were able to identify differences in model performance and when and where they occurred. A number of scientists have stated a need for more studies like this, i.e., evaluations of evaluation methods for diffusion models.

There is sometimes a need to adjust diffusion models for different sampling times. A number of authors have suggested power laws in sampling time to correct σ_y (Gifford, 1975). These are mostly derived from application of statistical theory to wind direction fluctuation measurements, but such corrections should be functions of stability and distance, i.e., existing methods tend to be overly generalized. Measurements and models appropriate to the χ' question will also probably serve to improve our handling of the sampling time question.

On a final note, a need frequently voiced that has much to do with research is to convincingly demonstrate how uncertainty in predictions of χ can be usefully incorporated in routine regulatory decision making.

STABLE BOUNDARY LAYER ISSUES

Next to χ' and model evaluation needs, the most frequently voiced research needs have to do with SBLs. Perhaps this is in part because there has been so much progress in the last decade on understanding diffusion in CBLs. The SBL has been relatively neglected partly because it is more difficult to make turbulence measurements in stable conditions, when turbulence intensities are quite low, and partly because SBLs are more complex than CBLs. They may have very large wind direction shears and are much more terrain sensitive. This is because downslope gravitational forces on cooled air near the surface compete with large scale pressure forces and momentum transferred from above in forcing the flow, even on very slight slopes. At the AMS workshop in 1977 it was flatly admitted that "in very stable conditions our present knowledge is in a very poor state" (Hanna et al., 1977). Since the participants summarized the problem so well, here is the rest of the quote:

> "Our observations refer almost totally to "ideal" (flat, smooth) sites, and we have good evidence that roughness and terrain slopes greatly affect diffusion in the stable boundary layer. It is recognized that turbulent mixing normally occurs only in a shallow layer near the ground in these conditions, but the thickness of this layer varies over a very wide range, from meters to hundreds of meters, depending strongly on the upper wind and on site topography. Whether this layer encompasses or falls short of a given source height obviously will have a critical effect on diffusion from the source. We also need to know much more about the frequency of occurrence of "turbulence episodes" that have been observed at night. These greatly increase the mixing height, evidently for significant periods of time such as several hours. Except for the urban nocturnal boundary layer, which is convective and somewhat more akin to the daytime boundary layer, we have only begun to formulate possible ways to predict the nocturnal mixing height. Measurements of this height, for instance using a remote sensor such as an acoustic sounder, at a variety of sites and under a variety of wind and cloudiness conditions would greatly improve our knowledge."

There has been some progress in understanding the turbulent structure of momentum-driven SBLs since 1977 (Nieuwstadt, 1984a, b); this was greatly helped by high quality meteorological measurements made on a 200-m mast at Cabauw, the Netherlands, and an acoustic sounder to measure the height of turbulent mixing, h (Driedonks et al., 1978). However, this was a very flat site, with no significant driving of the boundary layer by downslope forces (Nieuwstadt claims that, in contrast to "the" CBL, there are many different SBLs, depending on the balance of driving forces). Even with this ideal site, the investigators decided to exclude data from some time periods because of evidence of gravity waves; these are long-period waves (several minutes or more) that cause h to fluctuate up and down and also cause quantities like σ_w to increase with height. Normally σ_w decreases with height and near zero values as z approaches h. The waves are not "turbulence", cause no mixing, and could perhaps be filtered out, but because they muddy the analysis of an ideal SBL, these periods were discarded. The effects of gravity waves and ways to predict their occurrence in SBLs are other areas completely void of information. The need to study SBLs in rolling or hilly terrains is also very great, as no work has yet been done in this area.

The most urgent need is for observations of h in a variety of terrains under different meteorological conditions, so that ways to model it using near-surface measurements may be tested; h has a rough correlation with u* and L for flat sites, but we have no appropriate measurements for sloped sites and no theoretical reason to expect the same correlations to apply. This is a most important parameter because turbulence drops to zero at $z = h$; h acts as a "lid" on diffusion from sources below h, and there is no vertical diffusion for sources above h.

Lateral diffusion over averaging periods like an hour remains an enigma for stable conditions. Not only is it very sensitive to terrain, it is sensitive to the history of the flow for hours upwind. This is because when turbulence is suppressed due to static stability there is almost no dissipation of energy in large horizontal eddies, which are not suppressed by static stability. It is like someone let the brake off. This is especially true above h, where there is no turbulence. The horizontal eddies that cause plumes to meander may originate from flow around a mountain, or a thunderstorm a few 100 km upwind many hours earlier or from the remnants of convective eddies from the previous afternoon. Horizontal eddies may be present in a wide range of degrees on different nights at the same site, depending on the history of the flow. For a high source above h at night, there is no consequence at night because there is no downward diffusion. However, when the mixing depth builds up to the plume level in the morning, fumigation occurs; whether this causes high concentrations or not depends on the lateral diffusion that occurred during the night. Below h, large horizontal eddies are very sensitive to terrain. For this reason, Pasquill recommended no σ_y curves except for 3-min averages, and the 1977 AMS committee strongly urged σ_a measurements at the site in question (Hanna et al., 1977). Recently it was found that nighttime σ_v fluctuations on a tower correlated with minute pressure fluctuations at the ground (Zhou and Panofsky, 1983). With more investigation, this might turn out to be an effective basis for categorizing σ_y at night by using a surface measurement.

The need for testing the effects of wind direction shear on σ_y was already mentioned; these are especially large effects in SBLs. Furthermore, we have almost no grasp on the prediction of the rate of change of wind direction with height at night, even in idealized terrain. This is an area that might be helped by second-order closure boundary layer models, which have some demonstrated success at simulating SBL flow on a slope (Brost and Wyngaard, 1978).

Turbulence episodes or bursts have been mentioned as a phenomenon requiring attention for nocturnal diffusion. These are sudden jumps in h, accompanied by increased turbulence and downward mixing of warmer temperature, larger $\overline{u}$, and any pollutants that were "captured" by the increase in h. These jumps occur when the stability in the layer above h becomes marginal, with a Richardson number $Ri = (g/\theta)(\partial\theta/\partial z)/(\partial\overline{u}/\partial z)^2$ near a value of 0.5 or so. This can happen as the wind profile above h evolves slowly due to the turning of the earth (before the turbulent burst the layer is practically frictionless, so $\overline{u}$ responds only to inertia and the horizontal pressure gradient); it can also be caused by weakening of $\partial\theta/\partial z$ by more rapid radiative cooling from above. When Ri falls below the critical value 0.25 somewhere in the layer, turbulence develops spontaneously and can spread at least through all layers where $Ri < 0.5$. These events are extremely difficult to predict. For the source below the original height of h, more vertical dilution will occur when h jumps, so such events would serve to reduce surface concentrations. However, for the source above the original h, accumulated pollutant may suddenly mix to the surface in such an episode, somewhat like morning fumigation. (Evidence of such events have been seen in nighttime SO_2 monitoring records from near a power plant, but high concentrations occurring during such an event have yet to be observed.)

CONVECTIVE BOUNDARY LAYER ISSUES

Although knowledge of diffusion characteristics of CBLs has advanced more than that of SBLs, important research needs remain. These needs are primarily for more extended, three-dimensional diffusion measurements, studies of late afternoon transitions, and studies of cloud effects.

Except for surface measurements near a surface source, measurements of diffusion in the CBL to date are overwhelmingly from laboratory studies. There were surface arc tracer measurements at one distance at Cabauw and at three distances in Copenhagen for elevated SF_6 releases in rather weakly convective conditions. These satisfy only a small part of the need for field data for comparison with the laboratory measurements, which showed such unexpected vertical dispersion behaviors. Some CBL experimental needs put forward by Ching et al. (1983) were (1) surface χ and $\chi(z)$ profiles to $x \approx 6$ km for both elevated and surface-source releases, (2) complete concurrent meteorological measurements including z_i, (3) long averaging or release times of at least several hours, (4) early morning and late afternoon releases to study transition, and (5) studies of cloud effects. Participants at the U.S. EPA-AMS Workshop on Updating Applied Diffusion Models (Weil, 1985) also expressed the desirability of including a dense surface sampling network and using remote sensing for three-dimensional plume measurements. Some of these experimental needs are addressed by the CONDORS experiment in Section 3. But, the CONDORS experiment did not address the issues of diurnal transitions and cloud effects. This experiment was limited by (1) only a single surface sampling arc at $x \simeq 1.2$ km, (2) settling and deposition of some of the chaff plume, and (3) oil fog detection and mapping by lidar to only $x \simeq 2$ km or dimensionless $X \simeq 1.5$. The critical area of maximum surface impact for elevated oil fog releases was well covered by lidar, but it could not distinguish oil fog from background haze at much larger distances due to highly diffusive conditions and limits on the amount of oil fog that could be released (this technique might be doubly effective at a remote, very low turbidity site).

The midday CBL usually approximates a steady state, with slowly increasing z_i and nearly constant w^*, so it is rather well understood. Many modelers feel uneasy about the application of steady-state models to transition periods early and late in the daytime. The growth of z_i in the early morning, with entrainment of initially stable air aloft, has been measured and modeled with some success; inversion breakup fumigation caused by z_i growth has been studied in the laboratory (Deardorff and Willis, 1982) and in a field experiment (TVA, 1970); this study predates convective scaling, but only crude estimates of w^* and z_i could be made from the tabulated measurements). What needs study the most is the dying-out of convective turbulence in the late afternoon and the concurrent collapse of the mixing depth, which tends to leave material "frozen" at heights that it mixed up to early in the afternoon (Ching et al., 1983). Kaimal et al. (1982) suggest that z_i may collapse rather suddenly in the late day, but evidence is sketchy.

Cloud effects on turbulence and diffusion have received very little attention in spite of the fact that clouds occur rather commonly. There are many degrees and types of cloudiness and many questions. Scattered cumulus clouds often form at the top of vigorous thermals; these have ground shadows that create cool spots of low or reversed surface heat flux. Does this phenomenon alter the balance of updrafts and downdrafts and affect vertical diffusion? When these clouds develop significantly into stable air above z_i, they may "vent" pollutants that were drawn into the supporting thermals; this phenomenon of cloud venting needs more study (Ching et al., 1983).

When the sky becomes very overcast, solar heating of the ground is cut off and H^* becomes very small, zero, or slightly negative, cutting off the buoyant forces driving thermals from the surface. However, cooling can occur at the top of the boundary layer due to radiative cooling of cloud tops and cool outflows; cooling from above can cause an "upside-down" CBL driven from above. This may sustain convective turbulence late into the afternoon on days with afternoon cumulus development. There is a basic need to study CBL turbulence in cloudy and partly cloudy conditions, which is probably done best with turbulence-instrumented aircraft. Heat and moisture fluxes need to be measured just below the cloud bases, as well as close to the ground, so that we can find out if current CBL models will work for top-driven CBLs merely by turning them upside down and using $-H^*$ at $z \simeq z_i$ in place of H^* at the surface. Very likely, further model modifications will be needed.

A good start in this direction has been made by the University of Wisconsin Boundary Layer Research Team (Stull and Eloranta, 1984; Wilde et al., 1985; and Stull, 1985) who studied the CBL with cumulus over Oklahoma. Albrecht et al. (1985) have studied the marine ABL topped with stratocumulus over the eastern Pacific. This is an example of the ABL being strongly cooled from the top.

CALMS AND OTHER INCONVENIENT EVENTS

A perennially dodged question is, "how do you model calms?" Most models have $\chi \propto \overline{u}^{-1}$, so their predictions for $\overline{u} = 0$ are alarming. One psuedo-answer to the question is that the wind never really stops blowing; it is just too slow for instruments to respond. When this happens, use the Gaussian model with $\overline{u} = 1$ m/s. A second such answer is just like the first, except that $\overline{u} = 0.5$ m/s is the magic minimum wind speed that makes the model work (this answer doubles χ, of course). These responses seem less than scientific, at the least. Plumes have been observed to rise and fan out in all directions in stable, nearly calm situations and have been observed to loop in all directions in nearly calm, convective conditions; in fact, this condition sometimes gives the highest observed mean surface concentrations at power plants (Bowne, 1984: personal communication). A few physically-based models for diffusion in $\overline{u} = 0$ conditions do exist, but very little has been published. Recently, a model for zero and very low $\overline{u}$ diffusion in the CBL was proposed (Deardorff, 1984). Adequate testing of such models requires a dense sampler network close to the source occupying all compass directions or remote sensing of a plume near the source and some sensitive wind speed measurements (e.g., sonic anemometers).

Another "inconvenient event" for which no special modeling techniques exist is rain. No diffusion experiments have been carried out in rain, and some instruments perform well only in dry weather. Continuous SO_2 monitoring has been made at many power plants, but accompanied by only standard, rather crude meteorological measurements. Furthermore, SO_2 is scavenged by rain and snow, which adds a large complicative factor to the analysis. Examination of such monitoring records should help answer the question of whether diffusion during precipitation events is a worst case. Scavenging and wet deposition are very important questions here, but are beyond the scope of the present paper. When precipitation comes from stratus clouds, as is usual for mid-latitude winter storms, there is almost no positive or negative heat flux at the surface, so we usually assume that "neutral" diffusion applies in these conditions. To my knowledge, no one has tested this assumption. When precipitation is convective, as in thunderstorms and on-again-off-again showers, large, cool downdrafts are generated by evaporative cooling in the rain cells. This condition undoubtedly affects diffusion in a profound way, but how?

RESOURCE ISSUES

Efficient use of research funding is an issue that is raised frequently in the context of research needs. As this section has shown, it is easy to assemble a rather long list of areas in which our knowledge is deficient. These areas are not "ivory tower" concerns, but affect many practical decisions involving investments that, collectively, total billions of dollars. Too much of the diffusion modeling supporting such decisions has been ad hoc or highly extrapolative, lacking appropriate validation for the application. Usually this reflects lack of appropriate data, but too often insufficient use is made of existing data that could provide partial validation, at least. More encouragement of evaluation efforts is needed, either through allocation of funds for contracted work or through allocation of scientific staff time. These efforts must also include a careful analysis of the range of applicability of any new models as well as outlining the details of model operation.

The long list of research needs necessitates prioritization, as there are not enough resources (including scientists) to address all the needs. However, prioritization is not enough; there is also the strategic problem of how to make the most of resources. Given a certain level of funding, one could choose to direct it all towards the top priority need, or divide it among the top 10 or 20 priorities. The consensus among diffusion research scientists seems to be that more is accomplished by supporting a multiplicity of small research projects with diverse goals, carried out by individuals or small teams, than by supporting a large project with narrow goals.

There are needs for occasional "big experiments", but these need not be budget busters. The Prairie Grass experiment of 1956, the most classic example, was a multi-agency effort; four universities and 60 scientists and technicians participated. The recent CONDORS convective diffusion experiment was more modest, involving about 20 people during data collection and costing about 1/4 million dollars (a large fraction of the cost was for computer processing of the remote sensing signals); yet, it was a well-instrumented experiment that accomplished its basic goal: to define diffusion in three dimensions for elevated and surface sources for highly convective conditions. Similar experiments on about this scale or a little larger (with more surface sampling) run in neutral and stable conditions as well as convective conditions at several diverse sites would give a far better experimental base for testing models and for checking against laboratory modeling. Laboratory modeling is a very cost-effective tool for basic diffusion research as well as for complex flow situations. It cannot provide all the answers, but it can be used to explore many fundamental phenomena in a very controlled and systematic way. Numerical modeling, such as second-order closure, LES, and Monte Carlo models also can be very cost-effective tools for basic explorations, if the models are reasonably efficient. Furthermore, it would also be cost-effective to support analyses of many under-used past meteorological and diffusion measurements in the light of contemporary boundary layer concepts.

SECTION 7

IMPLEMENTATION OF IMPROVED MODELING PRACTICES

The technical transfer problem

There is a very large gap between diffusion modeling as practiced and the state of the science at the research level. The Pasquill curves are still the basis of most diffusion modeling in the United States; they were developed around 1958 for short-term averages from low sources, contained many ad hoc assumptions, and were supported in a rather spotty way by diffusion and wind direction variance measurements. More informed modelers prefer the Brookhaven curves for elevated sources, but these too were developed in the same era with limited diffusion measurements at one field site and with many questionable assumptions (like Gaussian vertical distribution and non-evaporation of oil fog droplets). Obviously, more field and laboratory measurements on diffusion have been made since then. Tremendous progress has also been made in understanding turbulence in the ABL and in successful mathematical modeling of many aspects of atmospheric diffusion. At any conference or workshop concerning diffusion issues, however, one can sense much frustration over this situation.

It seems obvious that not enough effort is being made to translate research into better practice. What are needed are major efforts to simplify research models sufficiently to make them viable as practical tools and to find adequate substitutes for the parameters that are difficult to measure.

The frequent lack of effective measurements is a real handicap in this endeavor. If we must restrict our input variables to the type of measurements routinely made at airports up to now, we can do little more than define Pasquill categories and make crude estimates of L on the basis of wind speed sampled for 1 min every 3 hr at an unusually smooth site. Of course, there is little possibility of significantly improved diffusion modeling using such restricted measurements. For this reason, diffusion scientists have repeatedly recommended effective on-site instrumentation and upgraded measurements at NWS stations (Hanna et al. 1977; Strimaitis et al., 1981).

ESTIMATING NEEDED PARAMETERS FROM MORE OBTAINABLE MEASUREMENTS

Except for modeling diffusion at short distances, < 1 km, from sources low in the ABL, the mixing height, z_i or h, must be measured or estimated to do any kind of diffusion modeling. One cannot model at a distance pretending that there is no lid on diffusion. For elevated sources at night, there is a good chance that there is no vertical diffusion at all if z_s is above the nocturnal mixing depth, h; however, if $z_s < h$, vertical mixing down to the ground occurs and the whereabouts of h becomes a critical question. Other quantities required depend on the modeling approach; these include D, σ_e, $\overline{u}$, u*, L, and w*, discussed in Section 6. All of these should be measured in any diffusion research experiment for maximum utility of the results, but direct measurement of many of these parameters is too difficult to be recommended in applied modeling. Scientists are aware of the problem and have proposed a number of schemes for estimating the difficult parameters from those more easily measured.

There are many ways to estimate the surface-layer parameters u* and L (or $H^* = u^{*3}/(0.4L)$, which also appears in the definition of w*). The profile method, using $\overline{u}(z)$ and $\theta(z)$ profiles from a small tower, is most often suggested (Briggs and McDonald, 1978; Nieuwstadt, 1978; Irwin and Binkowski, 1981; Berkowicz and Prahm, 1982). However, site uniformity and height of vegetation are major factors to be considered when placing instruments. Without profiles, a crude estimate of the surface roughness is needed with any of the methods; this estimate can be made from site inspection (Wieringa, 1980). Pasquill and Smith (1983) give nomograms for estimating u*, L, and H* from wind at a single height, $\overline{u}_1$, and potential temperature difference ($\Delta\theta$) between 0.2 z_1 and z_1. In convective conditions, H* is highly correlated with $\Delta\theta$ alone (Dyer, 1965). In general, u* and L can be determined from $\overline{u}$ at a single height if an estimate of H* is available. Methods for doing this by using cloudiness and some site-dependent parameters for surface moisture have been developed (van Ulden and Holtslag, 1983; Holtslag and van Ulden, 1983). An alternative method developed by Berkowicz and Prahm (1982) requires net radiation and humidity deficit near the surface to estimate H*; both are relatively easy measurements to obtain (humidity deficit relates directly to temperature and relative humidity). A similar method for estimating L by using $\overline{u}$ and net radiation or insolation was given by Briggs (1982a); it is very simple, but lacks consideration of the surface moisture.

The variances of lateral and vertical wind directions are not as difficult to measure as turbulent fluxes, and a good choice of instrumentation is available (Kaimal et al., 1984a). Lacking measurement, they can be estimated from $\sigma_a \approx \sigma_v/\overline{u}$ and $\sigma_e \approx \sigma_w/\overline{u}$ by using the relationships that have been developed for σ_v and σ_w, with one exception: it is not possible to generalize σ_v or σ_a for stable conditions because of strong site effects. The vertical velocity variance is a function of u* and z/h in stable conditions (Nieuwstadt, 1984a) and of u* and z/L in neutral and unstable conditions (Panofsky et al., 1977). In the daytime, σ_v is a function of u* and w* (Panofsky et al., 1977).

The most difficult parameters to estimate accurately are the mixing heights, z_i for the CBL and h for the SBL. Unfortunately, these are also the most important parameters for many applications. Direct measurements are best, and this will become increasingly affordable as better acoustic and laser sounders are produced in greater numbers. Kaimal et al. (1982) tested five remote sensing techniques for measuring z_i against tower, aircraft, and rawinsonde determinations. The authors also showed that z_i strongly correlated with the spectral peak frequency of wind speed near the ground, f_u, with $z_i \approx 0.9\ U/f_u$; U is the mid-mixing layer value of $\overline{u}$, which in the CBL follows surface-layer similarity relationships up to about 0.1 z_i, and then is essentially constant at the 0.1 z_i value (Holtslag, 1984). Lacking this type of measurement, a slab model for z_i can be applied that requires the time-integrated heat flux (or H*) since sunrise and an early morning temperature sounding such as those made routinely at major airports by the NWS. Several such models were tested against observations from 10 days at Cabauw in the Netherlands (Driedonks, 1982); excellent results were obtained from models that included a u* term in addition to the above inputs.

For the nocturnal h, Nieuwstadt and Tennekes (1981) developed a prognostic equation that tested well against observations at Cabauw, but it requires many input parameters (including the geostropic wind). Nieuwstadt (1984b) also compared some simpler schemes for h against Cabauw observations; he found fairly good correlations with Zilitinkevich's (1972) steady-state prediction, $h = 0.4\ (u^*L/f)^{1/2}$, and with a semi-empirical equation, $h = 28\ \overline{u}_{10}^{\,3/2}$ (mks units). The constant in the latter equation is dimensional, and should be regarded as site-dependent. It is very similar to the semi-empirical equation of Venkatram (1980), $h = 2400\ u^{*3/2}$ (mks units), which gave a good fit to Minnesota measurements (Caughey et al., 1979).

Except for σ_y in stable conditions, for which there is no substitute for measurements of σ_a, it appears that all meteorological parameters for modeling diffusion can be estimated from a morning rawinsonde sounding, $\overline{u}$ at a height of 10 m or so, and some type of sensible heating/cooling estimate; the latter requires solar radiation, net radiation, or cloudiness and solar elevation and possibly a humidity measurement.

What is greatly lacking at present is independent testing of these parameter substitutions at different sites. For almost every technique outlined above, testing was conducted at only a single site. There is no way of knowing whether the methods transfer to sites of different character.

Furthermore, in cases in which there are many approaches, as for the estimation of u* or L, we would like to know which one is best and which is the simplest technique that is "good enough."

All of these questions could be settled nicely with comparative meteorological measurement experiments carried out during some daytime and some nighttime periods at three or four contrasting sites, for instance, flat and rather smooth; suburban or forested; gently rolling; and hilly. Full meteorological measurements should be made concurrently, so that any simplified scheme can be tested against the direct measurements. Such a program would accelerate improved modeling efforts tremendously and save years of guesswork, non-optimum on-site measurements, and modeling attempts with the wrong input parameters.

BUILDING BETTER OPERATIONAL MODELS

Implementing improved models requires research and commitment. Adding an occasional experiment with partially complete meteorological measurements to the potpourri of past experiments and the continued development of scientific, albeit esoteric, models is not enough for progress in a practical sense if the last 20 years' experience is an indicator. Making the technical transfer happen requires a large effort to sort through the numerous scientific models (Section 5), determine the major features having the most practical significance, search for effective substitutes for the difficult-to-measure meteorological inputs, sift through the available diffusion data for experiments with the required measurements and no debilitating uncertainties, perform extensive comparisons between the data and several promising models plus the old models (both with and without various substitute meteorological parameters), evaluate the results for the most effective models given several different quality levels of input parameters, and then cast the favored model(s) into operational form with computer programs and/or nomograms. Most of this hard work would be classified as applied research and a prime research priority (the only research deserving higher priority is conducting experiments that are designed to facilitate meteorological and diffusion model comparisons and that make complete and trustworthy measurements).

The task so briefly stated above is so large, especially when the variety of operational modeling needs is considered, that no individual or small group could carry it out in just a few years, even if they were relieved of all other responsibilities. A few scientists have made efforts in this direction, but their time is divided among many competing tasks, so that they cannot tackle the whole of the technical transfer problem. Another factor for research scientists to consider is that there is more recognition for original research than for midwifing technical transfer. However, many scientists achieve a great sense of satisfaction when their research results are translated into something practical, and they are willing to assist in the process.

A few of the efforts towards better operational models should be mentioned. Pasquill's (1961) original Gaussian scheme was such an effort, on the part of a research scientist, and was a great step forward at the time. Smith (1973) refined this system for vertical diffusion from a source near the ground by introducing continuous, rather than discrete, stability categories and providing convenient nomograms, including one for a correction for surface roughness (his curves were based on the results of an eddy-diffusivity model). Irwin (1979) first advanced a generalized scheme that used convective scaling for daytime diffusion and on-site σ_a and σ_e data to characterize σ_y and σ_z and provided meteorological substitutes for nonavailable σ_a and σ_e measurements. This has evolved into a more complete approach that makes use of the latest advances in knowledge of ABL structure, especially for stable conditions (Sivertsen et al., 1985; Irwin et al., 1985). Weil and Brower (1982) developed a model for power plant applications that uses u* and w* to characterize σ_v and σ_w; these determine short-range dispersion in the usual way, based on statistical theory (e.g., $\sigma_y = \sigma_v t$). These forms for σ_y and σ_z are then modified by simple, empirical functions of x.

The willingness of scientists to seek model simplifications in order to encourage practical applications was amply demonstrated at the 1984 Workshop on Updating Applied Diffusion Models, cosponsored by the U.S. EPA and the AMS (Weil, 1985; Briggs, 1985b). This willingness should be further encouraged by a program of sustained and systematic efforts to translate research results into improved modeling practice. This task requires a thorough familiarity with models and the available data sets for validation.

SECTION 8

ASSESSMENT

The purpose of this paper is to review the current state of science concerning basic diffusion in the ABL and identify major research needs in this area. Basic diffusion in this paper means diffusion of conservative, non-buoyant substances over reasonably homogeneous, ordinary terrain. Thus, the scope here excludes consideration of buoyancy, deposition, chemical transformation, complex terrain, and shoreline flow issues. Even so, a vast number of research papers concern basic diffusion, and there are many complexities to consider as various forces and surface conditions come into play in the ABL. This assessment section will summarize the state of knowledge concerning the mechanisms driving turbulent diffusion in the ABL, the scope of experimental information that has been obtained, the theoretical and experimental basis for diffusion modeling as currently practiced, and the strengths and weaknesses of a half-dozen approaches to improved diffusion modeling. It will then set forth what most needs to be done to expand the experimental base for model validation and further develop the most useful modeling approaches. Finally, it is suggested that more effort is needed to translate research advances into modeling practice. Model simplifications should be guided by comprehensive evaluations using all reliable experimental evidence.

ATMOSPHERIC BOUNDARY LAYERS

Progress toward the understanding of turbulent and mean structures of ABLs accelerated after the 1968 Kansas surface-layer field experiment (Izumi, 1971), in which accurate profiles of potential temperature and wind speed components, including fast-response turbulence measurements, were made on a 32-m tower. This experiment, among others, provided verification of the Monin-Obukhov surface-layer similarity theory, (which is discussed in Section 2). This theory has been very successful in predicting profiles of mean quantities and vertical turbulence in the surface layer; it also works fairly well for the horizontal turbulence velocities in neutral conditions and the high-frequency portion of the horizontal turbulence velocity in stable conditions. It fails to correlate with low-frequency horizontal velocity changes in stable conditions, which have periods ranging from about ~ 1/4 to 4 hr, if they are present at all; it also does not correlate well with horizontal turbulent velocities in unstable conditions.

The situation with respect to CBLs vastly improved after the convective similarity scaling was introduced by Deardorff (1970a,b) and was tested in the 1973 Minnesota experiment (Izumi and Caughey, 1976), which included tower measurements to 32 m and extended turbulence and mean measurements to 1200 m by mounting instruments on a cable tethering a large balloon. Convective scaling is discussed in Section 2. Following the analysis of the Minnesota experiment in terms of convective scaling, for the first time, the behavior of the major parameters of turbulent structure was understood for the whole of the ABL. Further results and confirmation of earlier results for CBLs have been obtained at many sites now, mostly by research aircraft. One very significant finding that affects diffusion is that the mode of vertical velocity is negative, about -0.5w*, throughout the mid-mixing layer; this explains why the centerlines of elevated plumes have been observed to descend in laboratory, numerical, and some field experiments. Little work has been done on the effects of partial cloudiness on CBL structure, and no work has been done on "upside-down" CBLs driven by radiative cooling at the top of a fog or heavy overcast of clouds.

Progress has been made in understanding NBLs, but mostly by means of numerical models rather than by experiments. Perhaps this is because truly neutral ($H^*/u^{*3} \simeq 0$) conditions are rather infrequent, and overcast, windy conditions are not the best flying weather. There is ample information for the lower part of the NBL, for which it appears that surface-layer similarity works very well for the vertical component of turbulence and fairly well for the horizontal components; these may be influenced by surface roughness or slope inhomogeneities. Higher up, Coriolis acceleration affects NBL structure, including turbulence, and the Coriolis parameter, f, is important. Further details of NBLs are contained in Section 2. There is need for more experimental measurements in the upper part of the NBL.

Much progress has been made in understanding SBLs since the Cabauw experiments begun in 1977, but understanding is far from complete because of the varied nature of SBLs. They are strongly influenced by the slightest terrain slopes and are also affected by radiative cooling, which depends on humidity and CO_2 profiles. Also, they are subject to disruptive events like gravity waves and turbulent bursts, which are hard to forecast. Much has been learned about undisrupted, pressure gradient-driven SBLs in flat terrain. This work is discussed in Section 2.

The diurnal cycle of ABLs is basically well understood, except that there is not much experimental information on the transitional periods near sunrise and sunset. The least understood phenomenon is the collapse or running down of CBL turbulence in the late afternoon, which may even overlap the beginning of SBL development at the surface; this collapse begins as soon as H* turns negative, about 1 hr before sunset. The rate of CBL collapse may depend on humidity, which increases radiation flux divergence and hastens the development of stable stratification aloft.

DIFFUSION EXPERIMENTS

Several dozen field experiments have been conducted using tracers or oil fog for basic diffusion studies, but only a few of them included a full range of desirable meteorological measurements, and these experiments are quite limited in terms of distance and plume sampling. Until recent experiments, the tracers released were not very conservative, which confounds interpretation of their results for vertical diffusion and surface concentrations. Laboratory modeling using tanks, water channels, or wind tunnels has been used with success in modeling diffusion in all stabilities, but its potential has not been exploited very much. In total, we have a patchwork quilt of experimental information that contains many blank squares and many uncertain ones, especially for vertical concentration distributions and diffusion at distances greater than 3 km.

Almost all field experiment estimates of σ_z, the vertical diffusion parameter, were made using only surface concentration measurements by assuming a Gaussian vertical distribution of concentration, χ, and a conservative tracer, i.e., $\iint(\chi\overline{u})dydz = Q$, the source strength. It has gradually come to light that all the tracers used, except for recent gaseous tracers like SF_6, were not conservative; SO_2 gas and particulate tracers deposit significantly, and oil fog evaporates substantially within a travel distance of 2 km or less. Unless reasonably good estimates can be made of the deposition or tracer loss rates and the data can be appropriately reanalyzed (e.g., Gryning et al., 1983), the σ_z results from these experiments must be held questionable. The same is true of the crosswind-integrated concentration, $\int \chi dy$. We have reliable surface χ measurements for SF_6, but only for neutral and slightly unstable conditions at $x = 2$ to 6 km for elevated releases and only for very stable conditions at $x < 0.4$ km for surface releases. In a few of the earlier experiments, tower or balloon tethering cable measurements were obtained for vertical distributions from elevated releases before significant deposition could occur, but these were complete only for 26-m releases in neutral to stable conditions and ~ 200-m releases in near-neutral conditions.

For lateral diffusion, σ_y, deposition is less of a problem and there is a more complete collection of reliable tracer measurements; these include some to 26 km for surface releases and some to 10 km for elevated releases. The meteorological measurements accompanying the longer range experiments was sufficient for testing only certain classes of models. As we look at available plume width measurements at greater distances, up to 100 km or so, we find rather scant meteorological information. Many of these measurements are quasi-instantaneous plume widths obtained by a single aircraft traverse at a given distance. Under some circumstances, observations of buoyant plume parameters can be used to approximate passive plume dispersion, especially for σ_y after vertical mixing is complete, as in the CBL; this gives useable data in the 15- 30-km range for σ_y in unstable conditions.

Most field experiments suffered from meteorological instrumentation deficiencies, so they cannot be used to test all types of models. The earlier turbulence instrumentation was not accurate, and many experiments lacked sufficient wind and temperature profile measurements and even simple measures like solar insolation and cloudiness. Only one of the earliest experiments included soundings adequate for obtaining z_i, a key CBL parameter. The most recent field programs have included measurements of z_i and a complete array of good meteorological measurements. Hopefully, this is indicative of better experiments to come. Not only do we now have good turbulence instrumentation available, but we have much better tracers and remote plume sensors like lidar, which has been used in two very recent basic plume diffusion experiments.

Fluid modeling has proven a reliable tool for studying many aspects of diffusion in ABLs. It is cheaper than field experiments and is far more controllable, although the drastic reduction in scale does impose some limits on what can be simulated, particularly when buoyancy is involved. For instance, it is not possible in present facilities to simulate the effects of large horizontal eddies or wind direction shear on σ_y, but otherwise fluid modeling has proven quite versatile in simulating vertical turbulent diffusion and the minimum σ_y in the absence of the above effects.

A recent series of wind tunnel experiments at Colorado State University simulated basic diffusion for an elevated release and a surface release over a smooth surface and over a rough surface for CBLs, NBLs, and SBLs (Cermak et al., 1983). For the unstable runs, fairly good agreement was obtained with other CBL diffusion experiments, in spite of short dimensionless fetch and weak capping. Moderately stable ABLs ($h/L = 4$ to 6) were obtained by using a strong temperature differential, 78°C, between cooled floor and heated inflow, which is necessitated by the scale-down. Other wind tunnels have given excellent comparisons with field measurements in neutral conditions for σ_z as a function of x and z_0, the roughness length, but with σ_y about 20% smaller than in the field under similar conditions (this is perhaps due to the constraining effect of the tunnel walls). A similar decrease in σ_y, compared to field experiments interpreted in terms of convective scaling, has been noted for the convective tank experiments of Willis and Deardorff (1976, 1978, 1981). Their unique facility, basically just a water tank with a uniformly heated bottom, provided the pioneering results for diffusion in very convective conditions. The somewhat controversial results for vertical diffusion from this series of experiments have been essentially duplicated, by using convective scaling, in numerical experiments in the CSU wind tunnel and in the few field observations for which adequate measurements exist.

Both wind tunnels and convective water tanks could be put to further use to make systematic studies of source height effects, surface roughness effects, surface roughness and surface heat flux inhomogeneity effects, and concentration fluctuation statistics, a matter of great current interest. Also, as wind tunnels and water channels have been successfully used to simulate diffusion in complex terrains, it seems that they could also be used to study the effects of typical rolling or hilly terrains on diffusion.

DIFFUSION MODELING AS PRACTICED

The basic diffusion models used in standard practice today were developed 25 to 30 years ago. The stability classification schemes that they use, although crude, are not "unscientific". They have some basis in the physics relevant to ABL diffusion, but this basis is quite incomplete in the light of what has been learned about ABLs since 1968. The σ_y and σ_z curves that were empirically determined for each "stability class" were based largely on single-site data consisting of only surface measurements of χ for non-conservative tracers, a tiny fraction of the diffusion data now available. It is no wonder that scientists have been calling for a overhaul of the system for years now. All the early schemes also assumed a Gaussian-shaped plume distribution, both vertically and laterally, based on a crude analogy of turbulent diffusion with molecular diffusion.

This assumption, while far from rigorous, has proven amazingly resiliant. For sufficient averaging times, 1 hr or so, it works as well as any shape for lateral diffusion. Because vertical turbulence is not homogeneous, systematic deviations from the Gaussian shape have been observed for $X(z)$ profiles. In most cases, the consequences of using Gaussian $X(z)$ for ground-level concentration predictions are not great and the prevailing mood seems to be in favor of not changing. However, for vertical diffusion in unstable conditions in CBLs, the consequences appear to be considerable.

The Brookhaven stability classification scheme (Singer and Smith, 1966), based on wind direction fluctuations, was an adaptation of a scheme proposed at least as early as 1932 (Giblett, 1932) and founded on statistical theory (Taylor, 1921). The basic result for short-range diffusion, $\sigma_y = \sigma_a x$, where σ_a is the lateral wind direction variance, can be stated succinctly as "the plume goes as the wind blows" and is so direct as to defy any contrary argument (as long as the wind vane faithfully responds to the wind). The correlation between σ_y and σ_a becomes less direct in a few tens of meters of travel, and it was an intuitive assumption that some empirical correlation would remain at distances of tens of kilometers. This assumption has held up, although it can be improved upon. A more risky assumption was that vertical diffusion would correlate with σ_a also; this has proven true at short ranges in unstable and neutral conditions when turbulence eddies are more or less isotropic (high as the they are wide); however, it is definitely risky for stable conditions when site- and time-specific large horizontal eddies, or meanders, can make lateral wind fluctuations and dispersion much larger than vertical, at least over 1-hr averages. The Brookhaven curves were based mostly on surface measurements of oil fog droplets from a plume released at $z_s = 108$ m in rolling, forested terrain. The usual assumptions of Gaussian $\chi(z)$ and a conservative tracer were applied to infer σ_z values from surface concentrations. As even heavy-oil fogs can lose 50% of their lidar cross-section in about 2 km of travel, there is reason to doubt the conservative-tracer assumption and the resulting σ_z values. The surface σ_y values should not have been greatly affected by oil fog evaporation.

Pasquill's (1961) stability classification scheme, which was adopted by Gifford (1961) and adapted by Turner (1961) for use with Pasquill's curves, is based upon measurements usually available at airports, wind speed at $z \simeq 10$ m and insolation or cloudiness. It is an intuitive scheme, but as was shown later, it correlates significantly with the physically very important Obukhov length, L (Golder, 1972; Briggs, 1982). It only lacks adjustment for the effect of local surface roughness. Like L, it gives good correlation with vertical turbulence and diffusion near the surface. It correlates with lateral turbulence and σ_y only when the turbulence is nearly isotropic, as for 3-min averages in neutral and stable conditions. In unstable conditions, turbulence is somewhat isotropic at an elevation of $0.1\ z_i$ or more, but not at the ground; z_i has as much influence on σ_y as does L, and Pasquill categories have little to do with z_i. Pasquill's σ_y and σ_z curves were intended for use with near-surface sources, as they are based on near-surface data (also, the categorization scheme, like L, has decreasing relevance higher up in the ABL). The σ_y curves are based mostly on 3-min averaged wind direction fluctuation data, using statistical theory. They apply to 3-min averages, and give quite conservative estimates if applied to 1-hr average σ_y values. The σ_z curves are based mostly on an eddy-diffusivity theoretical prediction for the "neutral" category for a rather smooth surface (z_0 = 0.03 m), the effect of stability was determined and extrapolated from the Prairie Grass data, which extended only to 0.8 km. Deposition effects probably made these σ_z estimates too large, especially at larger x in stable conditions, and non-Gaussian $\chi(z)$ effects in very unstable conditions probably caused the σ_z estimates to grow too fast at $x > 400$ m, although the rapid drop in surface concentrations was quite consistent with Willis and Deardorff's results (Nieuwstadt, 1980).

Some brief comments are due on two other stability classification schemes. Hogstrom's (1964) suggestion of $s = \Delta\theta/u_f^2$ as an index in which u_f is the "free wind" velocity at about 500 m was based on testing of five different indices for correlation with observed σ_y and σ_z values from his oil fog puff experiments. For his site, u_f worked better than $\overline{u}$ at source height z = 87 m. The potential temperature difference was measured between 30 and 122 m, bracketing the source height. At least in the surface layer, $\Delta\theta/u^2$ is a bulk Richardson number that does relate directly to L. Using u_f and $\Delta\theta$ from higher up, there would not be a direct correspondence, but s would broadly indicate dynamic stability in the elevated layer encompassed by $\Delta\theta$. For Hogstrom's experiments, which were made only in neutral to very stable conditions, s correlated very well with changes in relative diffusion, σ_{yr} and σ_{zr}, and with the diffusion of the puff centerlines in the vertical, σ_{zc}. There was no discernible trend with stability in σ_{yc}, the meandering part of lateral diffusion. The successes, and failure, of this index for Hogstrom's elevated releases roughly parallel those of Pasquill's scheme, which is likewise loosely related to L. However, if Hogstrom had been able to use his puff technique in unstable conditions, he probably would not have found much correlation with s, as $\Delta\theta$ is very small in CBLs above 30 m (for good correlation with L, $\Delta\theta/\overline{u}^2$ should be measured below 20 m; the bottom level should be as low as 1 m, if practical). For the same reason, the use of ΔT at considerable elevation as a stability index could not distinguish at all between unstable and neutral conditions. Measured near the surface, ΔT or $\Delta\theta$ correlates well with H* in unstable conditions, but not with $\overline{u}$ or u*, which have much more effect on diffusion. Thus, ΔT alone is inferior as a stability index; even $\overline{u}$ alone broken into a "day" and a "night" category works much better (Briggs and McDonald, 1978).

ALTERNATIVE MODELING APPROACHES

Many papers have been published concerning more scientific approaches to basic diffusion modeling. Thankfully, they sometimes include testing of a model against some kind of diffusion observations. However, comprehensive testing of a variety of alternative approaches has not been done; meteorological measurements in data sets make such comparisons very difficult, if not impossible. This paper has divided the approaches into six generic categories for discussion and attempts to identify the strengths, weaknesses, and potentials of each method.

Surface-layer similarity theory expresses gradients of mean quantities in terms of $(u^*z)^{-1}$ times their surface fluxes times functions of z/L and expresses turbulence variances as u^* times functions of z/L. This theory has been quite successful at predicting profiles of $\overline{u}$, θ, and other mean quantities in the lower ABL. Near the surface, $\sigma_w/u^* \simeq 1.3$; in CBLs it is a function of z/L when $z < 0.1\ z_i$, but it is influenced by $z_i/|L|$ higher up. In SBLs, it is a function of z/h, and h and L are rather strongly correlated with each other. Thus, there is a broad region of validity for this theory for vertical turbulence. There have been ample demonstrations that it also works well for vertical diffusion, at least for surface sources when $\sigma_z < 0.3\ z_i$ or h. There is some evidence that this theory works well for minimum σ_y in neutral and stable conditions, as in wind tunnel tests, which lack the large horizontal eddies sometimes present in the atmosphere. The method gives relatively poor results for σ_y in unstable conditions, for which w^* has proven a more relevant velocity scale than u^*.

In unstable conditions, at least in CBLs with $z_i/|L| > 15$ or so, there is little doubt that convective scaling is a better approach. This replaces u^* and L with w^* and z_i as principal scales. Convective scaling has been demonstrated in numerous measurements programs to correlate well with turbulent velocities throughout CBLs (surface-layer similarity is superior only for σ_w below $z = |L|$). Experience has shown that approaches that order turbulence measurements well also order diffusion well. The fact that similar results are obtained for diffusion in such diverse media as water tanks, wind tunnels, numerical experiments, and field experiments when the results are nondimensionalized with convective scaling does much to instill confidence in the method. The correct scaling, in itself, does not constitute a diffusion model; it merely provides a way to summarize data efficiently and in a more universal format to allow comparisons of data from diverse experiments. It also suggests a simplified format for diffusion models. A number of equations for σ_y and σ_z and non-Gaussian vertical diffusion models using w* and z_i scaling have been proposed (Briggs, 1985b). Some of these also include neutral asymptotes that use u* as a second velocity scale.

Gradient-transfer, or "eddy diffusivity", models assume that the flux of a quantity is proportional to its gradient, and that the quantity moves toward the region of lower concentrations. The constant of proportionality, K, is called the eddy diffusivity in turbulent flow and has units of length^2/time. This theory is built on the same analogy to molecular diffusion as the Gaussian plume assumption and shares some of its virtues and weaknesses. It is rarely valid for lateral diffusion, as σ_y must be larger than the largest horizontal eddies for the assumption to hold true. In the CBL, $K_y \simeq 0.1 w^* z_i$ does provide a supportable asymptotic prediction for the minimim σ_y, $\simeq 0.5\ (w^* z_i x/U)^{1/2}$, but usually horizontal motions on a larger scale than z_i are present and an $x^{1/2}$ asymptote is not observed (Briggs, 1985b). Gradient transfer theory has the most validity for vertical diffusion from a surface source because the largest eddies affecting this diffusion are not much larger than σ_z (except in CBLs at $x > 0.5\ \overline{u}\ z_i/w^*$ when the mode of $\chi(z)$ lifts off the ground against the predictions of gradient-transfer models). It has been a useful tool for producing practical σ_z predictions when good estimates of K(z) are made by using surface-layer similarity scaling (Smith, 1973). Some more advanced models, like spectral diffusivity theory, attempt to account for non-local effects on fluxes of concentration; these have not been developed enough to encourage practical application.

A more advanced technique is second-order closure modeling, which assumes that third-moment quantities like $\overline{u'w'\chi'}$ are functions of local second-order or mean quantities or their gradients. Any local closure technique works best for stable conditions when turbulent eddies are small, and second-order closure models have been highly successful at predicting the behaviors of SBLs, including the effects of terrain slope and radiative cooling. They have hardly begun to be applied to turbulent diffusion in SBLs, but it seems that they would provide an excellent exploratory tool, especially for the effects of z_s/h and terrain slopes.

Large eddy simulation (LES) models were first applied to prediction of global weather patterns and have provided very good results for turbulent structure in CBLs and NBLs; both of these contain large eddies, of the order of their height. Unlike the second-order closure models, which predict time-averaged properties, the big-eddy models compute the instantaneous structure of eddies larger than the computational grid size. This provides actual "pictures" of the turbulence structure. To get averaged quantities, one computes the time averages by "sampling" over many time steps, much as in a field experiment. To account for the interactions between adjacent cells, an eddy-diffusivity assumption is applied only for small scales, the "subgrid" scales. This permits counter-gradient diffusion to occur on the large scales, which does occur in simulated CBLs. These models have given very good results for diffusion in CBLs. They have further potential for this case and probably would also be a good tool to explore the effects of z_s/z_i and $z_i f/u^*$ on diffusion in NBLs.

Statistical models in various forms have long been workhorses of diffusion modeling. They are based on the idea that if enough is known about the statistics of turbulent motions, both mean instantaneous and time-averaged diffusion of passive substances following those motions can be predicted. The most immediate results, like $\sigma_y = \sigma_v t = \sigma_a x$ near the source, are straightforward and consistent with diffusion observations. Difficulties arise when turbulence is inhomogeneous, as it always is in the vertical, and because Lagrangian velocity statistics (which follow the flow) are required, while most turbulence measurements are Eulerian, i.e., from fixed sensors. Hay and Pasquill (1959) made progress toward solving the latter problem by assuming similar forms for Eulerian and Lagrangian autocorrelation functions, but with different times scales t_E and t_L that have a constant ratio. This ratio $\beta = t_L/t_E$ is inversely proportional to the turbulence intensity. Many useful results have developed from this approach, including Pasquill's σ_y curves, which are based on wind direction statistics. This approach has also been used to estimate the effect of sampling time on σ_y. Other adaptations of statistical theory have been made for attempting to predict relative diffusion and vertical diffusion, which is accomplished by inhomogeneous turbulence. Recently, several simple models for diffusion from elevated sources in CBLs have used observed vertical velocity statistics at source heights to predict $\chi(z)$ versus distance essentially by assuming that t_L is large.

Since 1976, there has been a move to improve σ_y and σ_z predictions by going back to the basic statistical theory prediction that $\sigma_y/(\sigma_v t)$ and $\sigma_z/(\sigma_w t)$ are functions of t/t_L. The motivations are that (1) this is surely correct at small t/t_L, (2) this is simple, (3) the functions of t/t_L are weak, not strong, so their form and the estimates of t_L are not critical, and (4) we now know enough about ABL structures to be able to estimate σ_v and σ_w at source height in many ways, if they are not measured directly. A number of functional forms and ways to approximate t_L have been suggested, and much testing against data sets has been performed for this method. It seems a desirable approach for generalized applied diffusion modeling and, once optimized, should give adequate predictions except when the Gaussian plume assumption itself is inadequate, i.e., for vertical diffusion in the CBL.

A further development along the statistical line of approach, but with "artificial" statistics, has been the random perturbation/random force/Markovian random walk/Monte Carlo/Langevin equation methods that have been revived and extended in recent years. The basic idea is to compute the velocity of particles in a turbulent flow assuming for each particle the velocity of the previous time step reduced by whatever is the drop in the autocorrelation function for the time step Δt plus a random velocity perturbation. This is provided by a random number generator in a computer, of course, and is distributed in a way that maintains the proper average velocity statistics for the flow. Many results can be produced by mathematical manipulation of the basic equations. Monte Carlo diffusion simulations, using thousands of computed particle trajectories, have produced further results, such as peak-to-mean concentrations, relative diffusion, and relationships between t_L and t_E. Methods have been developed to allow application of these techniques to vertical diffusion i.e., with inhomogeneous turbulence. For instance, the technique has been applied to vertical diffusion in the CBL with good simulations of experimental results. However, such applications require quite a few assumptions about the turbulence statistics and particle behaviors at boundaries. It would seem that these methodologies need considerably more testing against research-grade observations, perhaps wind tunnel observations, before they can be relied upon as general purpose research tools.

NEEDS FOR DIFFUSION EXPERIMENTS

Without question, there is a broad consensus that we must broaden the experimental base if we are to make progress towards better diffusion modeling. The many flaws and omitted measurements in most past experiments make them impossible to use for even-handed comparisons with both old and new modeling approaches, although they are useful as auxiliary tests of certain models. The field experiments with adequate, or nearly adequate, meteorological measurements were too limited in distance range or stability range to provide for most of our model validation needs and usually had only surface concentration measurements. Loss of tracer by deposition was a problem in the great majority of experiments. Laboratory modeling of diffusion appears to be capable of fairly accurate simulations of most ABL diffusion aspects and could be used to a much greater degree to systematically explore the effects of source height and changed surface conditions.

The usefulness of any diffusion field experiment is determined by the completeness of the accompanying meteorological measurements as much as by the scope of concentration measurements. Most past experiments included only a selection of meteorological parameters, so they cannot be tested against most modeling approaches. A complete set of meteorological parameters includes (1) variances of vertical and lateral velocities or wind directions, σ_w and σ_v or σ_e and σ_a, (2) the surface flux quantities H^* and u^*, or Ri near the surface, or, at the least, $\Delta\theta/\overline{u}^2$ near the surface, (3) a measurement or estimate of the surface roughness, z_0, and (4) soundings to determine the capping height or mixing depth of the boundary layer, z_i or h. All of these are important for the testing of current modeling approaches such as statistical theory, surface-layer similarity theory, and convective similarity theory (the Obukhov length, $L = -u^{*3}/(kH^*)$, and the convective scale velocity, $w^* = (H^* z_i)^{1/3}$, can be derived from the above listing). This list also contains the most basic parameters for characterizing the turbulence structure in ABLs as far as we now know. In addition, it would be helpful to have complete profiles through the ABL or at least through the plume layer of $\overline{u}$, of θ (especially in stable conditions), and of mean wind direction, D (especially with measurements of σ_y at large x, where wind direction shear plays a role).

For experiments in stable conditions, the slope of the site is important, and in uneven areas the topography surrounding the site may be important. The site description should include inhomogeneities such as tree lines and varied surface cover. For the majority of past experiments, the Pasquill category cannot be determined because solar insolation was not measured and no notation was made of cloudiness. Such simple measurements should not be neglected. For alternative schemes, these measurements of net radiation and some notation or measure of surface humidity conditions should also be included.

In all conditions, plume concentration measurements are needed to larger distances. Some vertical profiles of concentration, even if very sparse horizontally, are greatly needed to at least the distance where the plume fills the boundary layer (from about 10 km in very convective conditions to perhaps 100 km in neutral conditions). Lateral plume width, or σ_y, measurements are needed to distances of the order of 100 km; even if measured at only a few distances, it seems that interpolation through large distance factors would be preferable to the present practice of extrapolating from 3 to 100 km. Three-dimensional plume concentrations can now be obtained by lidar within a few kilometers of the source, and excellent tracers with negligible deposition loss now exist that can be distinguished from background hundreds of kilometers away.

For stable conditions, in SBLs, the greatest need at present is for both boundary layer measurements and a few diffusion experiments at a variety of site types. Almost all that we know now pertains to flat sites or to sites with very slight slopes, ~0.1°; even this much slope has been observed to cause quite significant effects on h and on the turning of wind with height and time. Because of the great significance of h to diffusion modeling, it would be very helpful to have a "traveling boundary-layer experiment" (even without diffusion measurements) to measure h at sites of differing character - broad slopes, short slopes, hilly, rolling, etc. To correlate measured h with parameters measureable at the surface, we would also want u_*, L, $\overline{u}$, z_0, net radiation, cloudiness, and humidity. A portable instrumented tower, a tethersonde, and an acoustic sounder would be adequate basic tools for such experiments. We also need to develop some sort of climatology of nighttime σ_a at such sites, since it is so site-specific; this could easily be added to the same experiment. There is some evidence that minute pressure fluctuations at the surface correlate with large horizontal eddies or meanders aloft; measurements of these fluctuations would be useful. Apart from meander effects and wind direction shear effects, many aspects of SBL diffusion could be much more systematically and cheaply studied in a wind tunnel. These aspects include effects of source height, or z_s/h, z_0/h, and terrain undulations and surface roughness inhomogeneities.

For unstable conditions, in CBLs, our needs for three-dimensional plume information near the source are being partially met by the CONDORS experiment (Moninger et al., 1983), although it would be better to have that information extended to twice the distance to where uniform vertical mixing is nearly achieved. We still need some time-averaged σ_y measurements at $x \sim 100$ km, with CBL parameters measured; present observations extend to about 25 km. There is especially a need for turbulence measurements aloft in the late afternoon, as we have almost no information on CBL collapse; in addition to the usual CBL parameters, these measurements aloft should be accompanied by humidity profiles, as radiation flux divergence may greatly affect the process. Finally, as clouds are rather common in afternoon CBLs, we need to stop avoiding them and make turbulence measurements when clouds are present in varying degrees to see how correlations with CBL parameters are affected. In overcast, daytime conditions, there may be more cooling at the top of cloud layers than heating at the surface, and heat fluxes at the surface and at the cloud base may be in delicate balance. These conditions need to be measured at the same time as other CBL turbulence measurements. It is likely that some of these ABLs would turn out to be "neutral", which is good because there is so little information on the upper part of NBLs and the effect of $z_i f/u_*$ on their turbulent structure. These latter tasks could be performed by research aircraft.

There have been many calls for studies of concentration fluctuations, χ', and their statistics. These are needed for rational evaluation of modeling uncertainties, for determining sampling time effects, and for flamability and toxicity predictions for some gases. "Baseline" experiments on $\chi'/\overline{\chi}$ could most easily be performed in a wind tunnel, e.g., for an elevated and a surface source, over a rough and a smooth floor, and in unstable, neutral, and moderately stable conditions. For CBLs, such measurements could also be made in convective tanks, as some have already. In addition, because of uncertainties introduced by laboratory scale-down and the finite sample and probe sizes, some χ' measurements should be made in conjunction with full-scale diffusion experiments, even if only at a few locations per experiment.

Large voids exist in our understanding of and appropriate modeling of diffusion during stability transition, calms, and rain. It would be helpful, in the course of diffusion experiments in progress, to include a few runs that span the transition from unstable to stable conditions near sunset. Inversion breakup fumigation has received some laboratory tank study, but no full-scale experiment has been attempted with full CBL turbulence and profile development measurements. Calms are difficult to find in the atmosphere, especially during short periods of readiness to launch a field experiment, but they do occur. It seems likely that some case-study calm periods could be found among existing power plant SO_2 and meteorological monitoring archives; this, at least, is a place to begin. There is a real need to study calms, as near-zero wind periods during unstable conditions have produced "worst case" surface concentrations at some power plants. A study of this phenomenon could be made in a convective laboratory tank, but it probably would have to be two or three times wider than the one used up to now. As for rain, we have usually assumed that passive diffusion proceeds as it does in dry, neutral conditions, but there is no evidence that this is so; often, showery conditions are accompanied by convective downdrafts produced by evaporative cooling.

RESEARCH MODELING NEEDS

Because research diffusion models are numerous and seem somewhat ahead of the game at present in that there are generally inadequate diffusion observations for proper model validations, we have chosen to somewhat emphasize the experimental needs. We believe that this reflects the prevailing opinion in the scientific community at present. However, we do this with expectations that research model development will continue and will improve as better experimental data become available. Perhaps some approaches will stand out as more clearly rewarding when this happens. In the meantime, there are some areas in which we see more immediate benefits possible from some of the models.

We are beginning to understand the rudiments of SBL turbulence structure, but have not yet performed diffusion experiments concurrently with needed basic measurements like h. It seems that second-order closure modeling of SBLs so far has produced very good simulations of SBL evolution in time and of slope effects, at least on broad slopes. Surely these models could also be used to advantage to explore vertical diffusion in such boundary layers (these models presently only compute the vertical dimension). Slope effects, radiative cooling effects (probably small), and z_s/h effects could be readily explored, given adequate computer time. These models will also give the wind direction shear, but they cannot be used to predict lateral diffusion. For σ_y, further use could be made of statistical theories, including random perturbation modeling, especially with good wind direction fluctuation measurements.

For CBLs, it seems possible that radiative cooling could be added to a LES model. In conjunction with diminishing H* at the lower boundary, such a model could simulate CBL collapse, at least until the big eddies are no longer present. The overcast boundary layer, with various balances between surface H* and cloud base H*, could also be simulated with LES models. One advantage of these models is that they can be readily used to predict vertical and lateral diffusion from any height within the resolution of the grid cells; all it takes is sufficient computer time to compute numerous particle trajectories. A more speculative possibility for these models is simulation of turbulence and diffusion in CBLs topped by scattered clouds, which form at the top of more vigorous thermals; this would probably require some form of cloud updraft parameterization and some accounting of subsidence outside the clouds.

TECHNICAL TRANSFER NEEDS

Even without further research advances, we have gained so much new knowledge about boundary layer turbulence structure and diffusion since 1968 that it is greatly perplexing to many scientists that so little of this has filtered down to applied modeling. Of course, this is not an automatic process. Research models contain some parameters that are difficult to measure. The models are often complex entities that do not lend themselves well to routine usage, at least not in their original format. Furthermore, many operational modelers do not feel inclined to switch from the old schemes until they are proven inferior to an alternative scheme This requires, first of all, support of a scientific effort to validate, simplify, revalidate, and then package preferable models in a way that simplifies application and requires only minimal meteorological and source inputs.

In Section 7, we reviewed efforts that have been made to find adequate ways to estimate the required, difficult meteorological parameters by using simpler meteorological measurements. A rather large array of methods has been suggested, but, in most cases each method has only been tested with data from one site. This is inadequate, especially in stable conditions. One reason that full-meteorology diffusion experiments have been recommended here is that they allow any number of schemes for input simplification to be tested both against the fundamental ABL parameters, such as L, and diffusion data in a modeling context. Furthermore, any applicable model or model simplification can be tested against the same diffusion data. This is rarely achievable with past experiments, except for very recent ones. Only testing of old and new models side by side against "universal" data sets like the above will settle the question of how much modeling improvement, if any, can be obtained for various levels of effort.

Apart from diffusion experiments, traveling boundary-layer experiments like the one suggested here for SBLs would settle the question of which input simplification methods work best at sites of various characteristics. Any modeling approach, old or new, will work vastly better if we can make a better estimate of the turbulent mixing depth, h or z_i.

Such efforts will bear fruit, to the satisfaction of research scientists and model-users alike, only after some period of sustained support for researchers who can concentrate on the job of transforming validated research diffusion models into efficient and more accurate operational models than those in common use.

REFERENCES

Agterberg, R., F. T. M. Nieuwstadt, H. van Duuren, A. J. Hasselton, and G. D. Krijt, 1983: *Dispersion Experiments with Sulphur Hexafluoride from the 213m High Meteorological Mast at Cabauw in the Netherlands*. Royal Netherlands Meteorological Institute, DeBilt, The Netherlands, 129 pp.

Albrecht, B. A., R. S. Penc, and W. H. Shubert, 1985: An Observational Study of Cloud Topped Mixed Layers. *J. Atmos. Sci*., 42: 800-822.

Arya, S. P. S., 1981: Parameterizing the Height of the Nocturnal Boundary Layer. In: *Fifth Symposium on Turbulence, Diffusion, and Air Pollution*, Amer. Meteor. Soc. Boston, MA, pp. 111-112.

Baerentsen, J. H., and R. Berkowicz, 1984: Monte Carlo Simulation of Plume Dispersion in the Convective Boundary Layer. *Atmos. Environ*.,18: 701-712.

Barad, M. L., 1958: *Project Prairie Grass - A Field Program in Diffusion*, Volumes I and II. Geophysical Research Paper No. 59. Air Force Cambridge Research Center, Bedford, MA. [NTID PB151424 and PB151425].

Batchelor, G. K., 1949: Diffusion in a Field of Homogeneous Turbulence, I. Eulerian Analysis. *Aust. J. of Sci. Res*.,2: 437.

Batchelor, G. K., 1952: Diffusion in a Field of Homogeneous Turbulence, II. The Relative Motion of Particles. *Proc. Cambridge. Phil. Soc*., 48:345.

Berkowicz, R., and L. P. Prahm, 1980: On the Spectral Turbulent Diffusivity Theory for Homogeneous Turbulence. *J. of Fluid Mech*.,100: 433-448.

Berkowicz, R., and L. P. Prahm, 1982: Evaluation of the Profile Method for Estimation of Surface Fluxes of Momentum and Heat. *Atmos. Environ*. 16: 2809-2819.

Binkowski, F. S., 1983: A Simple Model for the Diurnal Variation of the Mixing Depth and Transport Flow. *Boundary-Layer Meteor.*, 27-236.

Bowne, N., 1960: Measurements of Atmospheric Diffusion From an Elevated Source. Proc. 6th A.E.C. Air Cleaning Conference, Atomic Energy Commission, 76-88. [NTIS TID-7593].

Briggs, G. A., 1981: Canopy Effects on Predicted Drainage Flow Characteristics and Comparisons with Observations. In: *Fifth Symposium on Turbulence, Diffusion and Air Pollution*, Amer. Meteor. Soc., Boston, MA, pp. 113-115.

Briggs, G. A., 1982a: Simple Substitutes for the Obukhov Length. *Third Joint Conference on Application of Air Pollution Meteorology*, Amer. Meteor. Soc., Boston, MA, pp. 68-71.

Briggs, G. A., 1982b: Similarity Forms for Ground-Source Surface-Layer Diffusion. *Boundary-Layer Meteor*.,23: 489-502.

Briggs, G. A., 1985a: Diffusion Modeling with Convective Scaling and Effects of Surface Inhomogeneities. Specialty Conference on Air Quality Modeling of the Urban Boundary Layer, Amer. Meteor. Soc., Boston, MA, (in press).

Briggs, G. A., 1985b: Analytical Parameterization of Diffusion: the Convective Boundary Layer. *J. of Climate and Appl. Meteor.* (in press).

Briggs, G. A., and K. R. McDonald, 1978: Prairie Grass Revisited: Optimum Indicators of Vertical Spread. *Proceedings of the Ninth International Technical Meeting on Air Pollution Modeling and its Application, No. 103, NATO/CCMS*, pp. 209-220.

Brost, R., and J. C. Wyngaard, 1978: A Model Study of the Stably-Stratified Planetary Boundary Layer. *J. Atmos. Sci*.,35: 1427-1440.

Businger, J. A., J. C. Wyngaard, Y. Izumi, and E. F. Bradley, 1971: Flux-profile Relationships in the Atmospheric Surface Layer. *J. of Atmos. Sci*., 28: 181-189.

Busse, F. H., 1970: Bounds for Turbulence Shear Flow. *J. Fluid Mech*. 41: 219-240.

Calder, K. L., 1949: Eddy Diffusion and Evaporation in Flow over Aerodynamically Smooth and Rough Surfaces: A Treatment Based on Laboratory Laws of Turbulent Flow with Special Reference to Conditions in the Lower Atmosphere. *Quart. J. Mech. and Appl. Math*.,II: 153.

Carpenter, S. B., T. L. Montgomery, J. M. Leavitt, W. C. Colbaugh, and F. W. Thomas, 1971: Principal Plume Dispersion Models: TVA Power Plants. *J. Air Pollut. Control Assoc.*, 21: 491-495.

Carras, J. N., and D. J. Williams, 1981: The Long Range Dispersion of a Plume from an Isolated Point Source. *Atmos. Environ.*, 15: 2205-2217.

Carras, J. N., and D. J. Williams, 1983: Observations of Vertical Plume Dispersion in the Convective Boundary Layer. *Sixth Symposium on Turbulence and Diffusion*, Amer. Meteor. Soc., Boston, MA, pp. 249-252.

Caughey, S. J., 1982: Observed Characteristics of the Atmospheric Boundary Layer. In: *Atmospheric Turbulence and Air Pollution Modeling*, F. T. M. Nieuwstadt and H. Van Dop, eds., D. Reidel Publishing Company, Dordrecht, Holland, pp. 107-158.

Caughey, S. J., and S. G. Palmer, 1979: Some Aspects of Turbulence Structure Through the Depth of the Convective Boundary Layer. *Quart. J. Roy. Meteor. Soc*.,105: 811-827.

Caughey, S. J., J. C. Wyngaard, and J. C. Kaimal, 1979: Turbulence in the Evolving Stable Boundary Layer., *Atmos. Sci*.,6: 1041-1052.

Cermak, J. E., 1963: Lagrangian Similarity Hypothesis Applied to Diffusion in Turbulent Shear Flow. *J. Fluid Mech.*,15: 49-64.

Cermak, J. E., P. K. Shirvastava, and M. Poreh, 1983: *Wind Tunnel Research on the Mechanics of Plumes in the Atmospheric Surface Layer: Part I.* (ER 83-84JEC-PKS-MP12), Colorado State University, Fort Collins, CO, 140 pp.

Chaudhry, F. H., and R. N. Meroney, 1973: Similarity Theory of Diffusion and the Observed Vertical Spread in the Diabatic Surface Layer. *Boundary-Layer Meteor*.,3: 405-415.

Ching, J. K. S., J. F. Clarke, J. S. Irwin and J. M. Godowitch, 1983: Relevance of Mixed Layer Scaling for Daytime Dispersion Based on RAPS and other Field Programs. *Atmos. Environ*.,17: 859-871.

Christensen, O., and L. P. Phahm, 1976: A Pseudospectral Model for Dispersion of Atmospheric Pollutants. *J. Appl. Meteor*.,15: 1284-1294.

Counihan, J. 1975: Adiabatic Atmospheric Boundary Layers: A Review and Analysis of Data from the Period 1880-1972. *Atmos. Environ*.,9: 871-905.

Cramer, H. E., 1957: A Practical Method of Estimating the Dispersal of Atmospheric Contaminants. *Proceedings of the Conference on Applied Meteorology*. Amer. Meteor. Soc., Boston, MA, pp. 33-55.

Cramer, H. E., 1976: Improved Techniques for Modeling the Dispersion of Tall Stack Plumes. *Proceedings of the 7th International Technical Meeting on Air Pollution Modeling and Its Application, No. 51*, NATO/CCMS, pp. 731-780.

Cramer, H. E., F. A. Record, and H. C. Vaughan, 1958: The Study of the Diffusion of Gases of Aerosols in the Lower Atmosphere. AFCRL-TR-58-239, MIT, 133 pp. [NTIS AD 152 582].

Davis, P. A., 1983: Markov Chain Simulations of Vertical Dispersion from Elevated Sources into the Neutral Planetary Boundary Layer. *Boundary-Layer Meteor.*, 26: 355-376.

Deardorff, J. W., 1970a: Preliminary Results from Numerical Integrations of the Unstable Boundary Layer. *J. Atmos. Sci*., 27: 1209-1211.

Deardorff, J. W., 1970b: Convective Velocity and Temperature Scales for the Unstable Boundary Layer for Rayleigh Convection. *J. Atmos. Sci*., 27: 1211-1213.

Deardorff, J. W., 1972: Numerical Investigation of Neutral and Unstable Planetary Boundary Layers. *J. Atmos. Sci*., 29: 91-115.

Deardorff, J. W., 1973: Three Dimensional Numerical Study of Turbulence in an Entraining Mixed Layer. Boundary-Layer Meteor. 1: 169-196.

Deardorff, J. W., 1974: Three Dimensional Numerical Study of the Height and Mean Structure of a Heated Planetary Boundary Layer. Boundary-Layer Meteor.,7: 81-106.

Deardorff, J. W., 1984: Upstream Diffusion in the Convective Boundary Layer with Weak or Zero Mean Wind. Fourth Joint Conference on Applications of Air Pollution Meteorology, Amer. Meteor. Soc., Boston, MA, pp. 4-7.

Deardorff, J. W., and G. E. Willis, 1974: Computer and Laboratory Modeling of the Vertical Diffusion of Nonbuoyant Particles in the Mixed Layer. In: Advances in Geophysics, 18B, H. E. Landsberg and J. van Mieghem, eds., Academic Press, New York, pp. 187-200.

Deardorff, J. W., and G. E. Willis, 1975: A Parameterization of Diffusion into the Mixed Layer. J. Appl. Meteor., 14: 1451-1458.

Deardorff, J. W., and G. E. Willis, 1982: Ground Level Concentration Due to Fumigation into an Entraining Mixed Layer. Atmos. Environ., 16: 1159-1170.

Deardorff, J. W., and G. E. Willis, 1984: Ground-level Concentration Fluctuations from a Buoyant and a Non-Buoyant Source within a Laboratory Convectively Mixed Layer. Atmos. Environ., 18: 1297-1309.

Deardorff, J. W., and G. E. Willis, 1985: Further Results from a Laboratory Model of the Convective Planetary Boundary Layer. Boundary-Layer Meteor., 32: 205-236.

DeBaas, A. F., and A. G. M. Driedonks, 1985: Internal Gravity Waves in a Stably Stratified Boundary Layer. Boundary-Layer Meteor., 31: 303-323.

Doran, J. C., T. W. Horst, and P. W. Nickola, 1978a: Variations in Measured Values of Lateral Diffusion Parameters. J. Appl. Meteor., 17: 825-831.

Draxler, R. R., 1976: Determination of Atmospheric Diffusion Parameters. Atmos. Environ.,10: 99-105.

Draxler, R. R., 1984: Diffusion and Transport Experiments. In: Atmospheric Science and Power Production, Darryl Randerson, ed. U.S. Dept. of Energy, NTIS No. DE84005177, pp. 367-422.

Driedonks, A. B. M., 1982: Models and Observations of the Growth of the Atmospheric Boundary Layer. Boundary-Layer Meteor.,23: 283-306.

Driedonks, A. G. M., H. Van Dop, and W. Kohsiek, 1978: Meteorological Observations on the 213m Mast at Cabauw in The Netherlands. *Fourth Symposium on Meteorological Observation and Instrumentation.* Amer. Meteor. Soc., Boston, MA, pp. 41-46.

Druilhet, A., J. P. Frangi, D. Guedalia, and J. Fontan, 1983: Experimental Studies of the Turbulence Structure Parameters of the Convective Boundary Layer. *J. Climate Appl. Meteor.*,22: 594-608.

Dyer, A. J., 1965: The Flux-Gradient Relation for Turbulent Heat Transfer in the Lower Atmosphere. *Quart. J. Roy. Meteor. Soc.*, 91: 151-157.

Eberhard, W. L., W. R. Moninger, T. Uttal, S. W. Troxel, and J. E. Gaynor, and G. A. Briggs, 1985: Field Measurements in Three Dimensions of Plume Dispersion in the Highly Convective Boundary Layer. *Seventh Symposium on Turbulence and Diffusion*. Amer. Meteor. Soc., Boston, MA, (in press).

Einstein, A., 1905: Uber die van Molekular Kinetischen Theorie der Warme Geforddr Bewegung von in Ruhenden Flussigkeiten Suspendierten Teilchen. *Ann. Phys*. 17: 549.

Elderkin, C. E., W. T. Hinds, and N. E. Nutley, 1963: Dispersion from Elevated Sources. Hanford Radiological Sciences Research and Development Annual Report for 1963, 1.29-1.36. [NTIS HW-81746].

Ellison, T. H., and J. S. Turner, 1959: Turbulent Entrainment in Stratified Flows. *J. Fluid Mech*.,6: 423-448.

Ferber, G. J., and R. J. List, 1973: *Long-Range Transport and Diffusion Experiments*, NOAA Air Resources Laboratories, Silver Spring, MD, unpublished.

Fiedler, B. H., 1984: An Integral Closure Model for the Vertical Turbulent Flux of a Scalar in a Mixed Layer. *J. Atmos. Sci*.,41: 674-680.

Fox, D. G., 1981: Judging Air Quality Model Performance. Summary of the Amer. Meteor. Soc. Workshop on Dispersion Model Performance. *Bull. Amer. Meteor. Soc*., 62: 599-608.

Fox, D. G., D. Randerson, M. E. Smith, F. D. White, and J. C. Wyngaard, 1983: *Synthesis of the Rural Model Reviews.* Environmental Sciences Research Laboratory, U.S. Environmental Protection Agency, Research Triangle Park, NC. 42 pp.

Fuquay, J. J., C. L. Simpson, and W. T. Hinds, 1964: Prediction of Environmental Exposures from Sources near the Ground Based on Hanford Experimental Data. *J. Appl. Meteor.*, 3, 761-770.

Garratt, J. R., 1982: Observations in the Nocturnal Boundary Layer. *Boundary-Layer Meteor*., 22: 21-48.

Garratt, J. R., and R. A. Brost, 1981: Radiative Cooling Effects within and above the Nocturnal Boundary Layer. J. Atmos. Sci.,38: 2730-2746.

Giblett, M. A., et al., 1932: The Structure of Wind over Level Country. Meteorological Office Geophysical Memoirs No. 54.

Gifford, F. A., 1961: Use of Routine Meteorological Observations for Estimating Atmospheric Dispersion. Nucl. Safety,2: 47-51.

Gifford, F. A., 1962: Diffusion in the Diabatic Surface Layer. J. Geophys. Res.,67:3207.

Gifford, F. A., 1975: Atmospheric Dispersion Models for Environment Pollution Applications. In: Lectures on Air Pollution and Environment Impact Analysis. Amer. Meteor. Soc., Boston, MA, pp. 35-58.

Gifford, F. A., 1976: Turbulent Diffusion Typing Schemes: a Review. Nucl. Safety.,17: 68-86.

Gifford, F. A., 1982: Horizontal Diffusion in the Atmosphere: A Lagrangian-Dynamical Theory. Atmos. Environ.,16: 505-512.

Gifford, F. A., 1983: Atmospheric Diffusion in the Mesoscale Range: The Evidence of Recent Plume Width Observations. Sixth Symposium on Turbulence and Diffusion. Amer. Meteor. Soc. Boston, MA, pp. 300-304.

Golder, D., 1972: Relations among Stability Parameters in the Surface Layer. Boundary-Layer Meteor.,3: 47-58.

Gryning, S. E., 1981: Elevated Source SF_6-Tracer Dispersion Experiments in the Copenhagen Area. RISO-R-446. Riso National Laboratory, Denmark.

Gryning, S. E., and E. Lyck, 1984: Atmospheric Dispersion from Elevated Sources in an Urban Area: Comparison between Tracer Experiments and Model Calculations. J. Climate Appl. Meteor.,23: 651-660.

Gryning, S. E., P. Van Ulden, and S. E. Larson, 1983: Dispersion from a Continuous Ground-Level Source Investigated by a K Model. Quart. J. Roy. Meteor. Soc.,109: 355-364.

Hanna, S. R., 1981: Turbulent Energy and Lagrangian Time Scales in the Planetary Boundary Layer. Fifth Symposium on Turbulence, Diffusion, and Air Pollution. Amer. Meteor. Soc., Boston, MA, pp. 61-62.

Hanna, S. R., 1983: Lateral Turbulence Intensity and Plume Meandering During Stable Conditions. J. Climate Appl. Meteor.,22: 1424-1430.

Hanna, S. R., G. A. Briggs, J. Deardorff, B. A. Egan, F. A. Gifford, and F. Pasquill, 1977: Meeting Review. AMS Workshop on Stablility Classification Schemes and Sigma Curves - Summary of Recommendations. Bull. Amer. Meteor. Soc.,58: 1305-1309.

Hanna, S. R., G. A. Briggs, and R. P. Hosker, Jr., 1982: Handbook on Atmospheric Diffusion. U.S. Dept. of Energy, NTIS No. DE82002045, 102 pp.

Haugen, D. A., Ed., 1959: Project Prairie Grass, a Field Program in Diffusion. Geophys. Res. Pap., No. 59, Vol. III, Rep. AFCRC-TR-58-235, Air Force Cambridge Research Center, 673 pp. [NTIS PB 161 101].

Haugen, D. A., and J. J. Fuquay, Eds., 1963: The Ocean Breeze and Dry Gulch Diffusion Programs, Vol. I. Air Force Cambridge Research Laboratories and Hanford Atomic Products Operations, HW-78435, 240 pp. (Available from Technical Information Center, P.O. Box 62, Oak Ridge, TN.)

Hay, J. S., and F. Pasquill, 1957: Diffusion from a Fixed Source from a Height of a Few Hundred Feet in the Atmosphere. J. Fluid Mech.,2: 299.

Hay, J. S., and F. Pasquill, 1959: Diffusion from a Continuous Source in Relation to the Spectrum and Scale of Turbulence. Atmospheric Diffusion and Air Pollution, F. N. Frenkiel and P. A. Sheppard, eds.,Advances in Geophysics, 6, Academic Press, NY, 345-365.

Hicks, B. B., 1985: Behavior of Turbulent Statistics in the Convective Boundary Layer. J. Climate and Appl. Meteor., 24: 607-614.

Hilst, G. R., and C. L. Simpson, 1958: Observations of Vertical Diffusion Rates in Stable Atmospheres. J. Meteor., 15: 125-126.

Hino, M., 1968: Maximum Ground-Level Concentration and Sampling Time. Atmos. Environ.,2: 149-165.

Hoffnagle, G. F., M. E. Smith, T. V. Crawford, and T. J. Lockhart, 1981: On-site Meteorological Instrumentation Requirements to Characterize Diffusion from Point Sources - A Workshop. Fifth Symposium on Turbulence, Diffusion, and Air Pollution. Amer. Meteor. Soc., Boston, MA, pp. 10-11.

Högström, U., 1964: An Experimental Study on Atmospheric Diffusion. Tellus,16: 205-251.

Högström, U., 1985: Von Karman's Constant in Atmospheric Boundary Layer Flow: Reevaluated. J. Atmos. Sci.,42: 263-270.

Holtslag, A. A. M., and A. P. Van Ulden, 1983: A Simple Scheme for Daytime Estimates of the Surface Fluxes from Routine Weather Data. J. Climate Appl. Meteor.,22: 517-529.

Holtslag, A. A. M., 1984b: Estimates of Diabatic Wind Speed Profiles from Near Surface Weather Observations. Boundary-Layer Meteor.,29: 225-250.

Horst, T. W., 1979: Lagrangian Similarity Modeling of Vertical Diffusion from a Ground-Level Source. J. Appl. Meteor.,18: 733-740.

Horst, T. W., J. C. Doran, and P. W. Nickola, 1979: Evaluation of Empirical Atmospheric Diffusion Data. U.S. Nuclear Regulatory Commission. NUREG/CR-0798. 137 pp.

Hosker, R. P., Jr., 1984: Flow and Diffusion Near Obstacles. In: Atmospheric Science and Power Production. Darryl Randerson, ed. U.S. Dept. of Energy, NTIS No. DE84005177, pp. 241-326.

Hunt, J. C. R., 1982: Diffusion in the Stable Boundary Layer. In: Atmospheric Turbulence and Air Pollution Modeling, F. T. M. Nieuwstadt and H. Van Dop, eds., D. Reidel Publishing Company, Dordrecht, Holland, pp. 231-274.

Hunt, J. C. R., J. C. Kaimal, and J. E. Gaynor, 1985: Some Observations of Turbulence Structure in Stable Layers. Quart. J. Roy. Meteor. Soc., 111: 793-815.

Hunt, J. C. R., and A. H. Weber, 1979: A Lagrangian Statistical Analysis of Diffusion from a Ground-Level Source in a Turbulent Boundary Layer. Quart. J. Roy. Meteor. Soc.,105: 423-443.

Irwin, J. S., 1979: Estimating Plume Dispersion - A Recommended Generalized Scheme. Fourth Symposium on Turbulence, Diffusion and Air Pollution. Amer. Meteor. Soc., Boston, MA, pp. 62-69.

Irwin, J. S., 1983: Estimating Plume Dispersion - A Comparison of Several Sigma Schemes. J. Climate and Appl. Meteor.,22: 92-114.

Irwin, J. S., and F. S. Binkowski, 1981: Estimating of the Monin-Obukhov Scaling Length Using On-Site Instrumentation. Atmos. Environ., 15: 1091-1094.

Irwin, J. S., S. E. Gryning, A. A. M. Holtslag, and B. Sivertsen, 1985: Atmospheric Diffusion Modeling Based on Boundary Layer Parameterization. EPA 600/3-85/056, Atmospheric Sciences Research Laboratory, U. S. Environmental Protection Agency, Research Triangle Park, NC, 43 pp.

Irwin, J. S., and M. Smith, 1984: Potentially Useful Additions to the Rural Model Performance Evaluation. Bull. Amer. Meteor. Soc.,65: 559-568.

Irwin, J. S., 1984: Site to Site Variation in Performance of Dispersion Parameter Estimation Schemes. In: Air Pollution Modeling and Its Application III, C. De Wispelaere, ed., Plenum, New York, NY, pp. 605-617.

Islitzer, N. F., 1981: Short-range Atmospheric Dispersion Measurements from an Elevated Source. J. Meteor., 18, 443-450.

Islitzer, N. F., and R. K. Dumbauld, 1963: Atmospheric Diffusion-Deposition Studies over Flat Terrain. Int. J. Air Water Pollut.,7: 999-1022.

Izumi, Y., and S. J. Caughey, 1976: Minnesota 1973 Atmospheric Boundary Layer Experiment Data Report. Air Force Cambridge Research Laboratory, Bedford, MA. 28 pp.

Johnson, W. B., 1982: Meteorological Tracer Techniques for Parameterizing Atmospheric Dispersion. Workshop on the Parameterization of Mixed Layer Diffusion. 20-23 October 1981, Las Cruces, NM, pp. 123-132.

Kaimal, J. C., N. L. Abshire, R. B. Chadwick, M. T. Decker, W. H. Hooke, R. A. Kropfli, W. D. Neff, and F. Pasqualicci, 1982: Estimating the Depth of the Daytime Convective Boundary Layer. J. Appl. Meteor.,21: 1123-1129.

Kaimal, J. C., J. C. Wyngaard, D. A. Haugen, O. R. Cote, Y. Izumi, S. J. Caughey and C. J. Readings, 1976: Turbulence Structure in the Convective Boundary Layer. J. Atmos. Sci.,33: 2152-2169.

Kaimal, J. C., J. C. Wyngaard, Y. Izumi, and O. R. Cote, 1972: Spectral Characteristics of Surface Layer Turbulence. Quart. J. Roy. Meteor. Soc.,98: 563-586.

Kaimal, J. C., J. E. Gaynor, P. L. Finkelstein, M. E. Graves, and T. J. Lockhart, 1984a: A Field Comparison of In-Situ Meteorological Sensors. Boulder Atmospheric Observatory Report Number Six. NOAA/ERL, Wave Propagation Laboratory, Boulder, CO, 96 pp.

Kaimal, J. C., J. E. Gaynor, P. L. Finkelstein, M. E. Graves, and T. J. Lockhart, 1984b: An Evaluation of Wind Measurements by Four Doppler Sodars. Boulder Atmospheric Observatory Report Number Five. NOAA/ERL, Wave Propagation Laboratory, Boulder, CO, 110 pp.

Kantha, L. H., and R. R. Long, 1980: Turbulent Mixing with Stabilizing Surface Buoyancy Flux. Phys. Fluids,23: 2142-2143.

Krishna, J. and S. P. S. Arya, 1981: Wind Structure in Neutral, Entraining PBL Capped by a Low-Level Inversion. Fifth Symposium on Turbulence, Diffusion, and Air Pollution. Amer. Meteor. Soc., Boston, MA, pp. 65-66.

Lamb, R. G., 1978: A Numerical Simulation of Dispersion from an Elevated Point Source in the Convective Planetary Boundary Layer. Atmos. Environ., 12: 1297-1304.

Lamb, R. G., 1979: The Effects of Release Height on Material Dispersion in the Convective Planetary Boundary Layer. Fourth Symposium on Turbulence, Diffusion and Air Pollution. Amer. Meteor. Soc., Boston, MA, pp. 27-33.

Lamb, R. G., 1982: Diffusion in the Convective Boundary Layer. In: Atmospheric Turbulence and Air Pollution Modeling. F. T. M. Nieuwstadt and H. van Dop, eds., D. Reidel Publishing Company, Dordrecht, Holland, pp. 159-230.

Lavery, T. F., A. Bass, D. G. Strimaitis, A. Venkatram, B. R. Greene, P. J. Drivas, and B. Egan, 1982: *EPA Complex Terrain Model Development Program: First Milestone Report - 1981*. EPA-600/3-82-036, Environmental Protection Agency, Atmospheric Sciences Research Laboratory, Research Triangle Park, NC, 304 pp.

Lee, J. T., and G. L. Stone, 1983a: The Use of Eulerian Initial Conditions in a Lagrangian Model of Turbulent Diffusion. *Atmos. Environ*., 17: 2477-2481.

Lee, J. T., and G. L. Stone, 1983b: Eulerian-Lagrangian Relationships in Monte Carlo Simulations of Turbulent Diffusion. *Atmos. Environ*., 17: 2483-2487.

Lewellen, W. S., and M. Teske, 1976: Second-Order Closure Modeling of Diffusion in the Atmospheric Boundary Layer. *Boundary-Layer Meteor*., 10: 69-90.

Ley, A. J., and D. J. Thomson, 1983: A Random Walk Model of Dispersion in the Diabatic Surface Layer. *Quart. J. Roy. Meteor. Soc*., 109: 847-880.

Li, W. W., and R. N. Meroney, 1984: Estimation of Lagrangian Time Scales from Laboratory Measurements of Lateral Dispersion. *Atmos. Environ*., 18: 1601-1611.

Lilly, D. K., 1967: The Representation of Small-Scale Turbulence in Numerical Simulation Experiments. *Proceedings, IBM Sci. Comput. Symp. Environmental Sci.*, IBM Form No. 320-1951, pp. 195-210.

Lu, N., W. D. Neff, and J. C. Kaimal, 1983: Wave and Turbulence Structure in a Disturbed Nocturnal Inversion. *Boundary-Layer Meteor*., 26: 141-155.

Lumley, J. L., and H. A. Panofsky, 1964: *The Structure of Atmospheric Turbulence.* John Wiley and Sons, Inc., New York, NY, 239 pp.

Mahrt, L., and R. C. Heald, 1979: Comments on "Determining Height of the Nocturnal Boundary Layer. *J. Appl. Meteor*.,18: 383.

Mahrt, L., J. C. Andre, and R. C. Heald, 1982: On the Depth of the Nocturnal Boundary Layer. *J. Appl. Meteor*.,21: 90-92.

McElroy, J. L., 1969: A Comparative Study of Urban and Rural Dispersion. *J. Appl. Meteor*., 8: 19-31.

McElroy, J. L., and F. Pooler, 1968: *St. Louis Dispersion Study, Volume II, Analysis*. AP-53, U.S. Public Health Service, National Air Pollution Control Administration, Cincinati, Ohio, 51 pp.

Monin, A. S., 1959: Smoke Propagation in the Surface Layer of the Atmosphere. In: *Advances in Geophysics*, 6, F. N. Frenkiel and P. A. Sheppard, eds., Academic Press, New York, pp. 331-343.

Monin, A. S., and A. M. Yaglom, 1971: Statistical Fluid Mechanics: Mechanics of Turbulence, Vol. 1, MIT Press, Cambridge, MA, 769 pp.

Monin, A. S., and A. M. Obukhov, 1954: Basic Laws of Turbulent Mixing in the Atmosphere Near the Ground. Tr. AK ad. Nauk. SSSR Geofiz. Inst. No. 24 (151): 163-187.

Moninger, W. R., W. L. Eberhard, G. A. Briggs, R. A. Kropfli, and J. C. Kaimal, 1983: Simultaneous Radar and Lidar Observations of Plumes from Continuous Point Sources. 21st Radar Meteorology Conference, Amer. Meteor. Soc., Boston, MA, pp. 246-250.

Nappo, C. J., 1984: Turbulence and Dispersion Parameters Derived from Smoke-Plume Photoanalysis. Atmos. Environ.,18: 299-306.

Nickola, P. W., 1977: The Hanford 67-Series: A Volume of Atmospheric Field Diffusion Measurements. Battelle Pacific Northwest Laboratories, 454 pp. [NTIS PNL-2433].

Nieuwstadt, F. T. M., 1978: The Computation of the Friction Velocity u* and the Temperature Scale T* From Temperature and Wind Velocity Profiles by Least-Square Methods. Boundary-Layer Meteor.,14: 235-246.

Nieuwstadt, F. T. M., 1980: Application of Mixed-Layer Similarity to the Observed Dispersion from a Ground-Level Source. J. Appl. Meteor., 19: 157-162.

Nieuwstadt, F. T. M., 1984a: Some Aspects of the Turbulent Stable Boundary Layer. Boundary-Layer Meteor.,30: 31-55.

Nieuwstadt, F. T. M., 1984b: The Turbulent Structure of the Stable Nocturnal Boundary Layer. J. Atmos. Sci.,41: 2202-2216.

Nieuwstadt, F. T. M., and H. Van Duuren, 1979: Dispersion Experiments with SF_6 from the 213m High Meteorological Mast at Cabauw in The Netherlands. Fourth Symposium on Turbulence, Diffusion and Air Pollution, Amer. Meteor. Soc., Boston, MA, pp. 34-40.

Nieuwstadt, F. T. M., and H. Tennekes, 1981: A Rate Equation for the Nocturnal Boundary-Layer Height. J. Atmos. Sci.,38: 1418-1428.

Nieuwstadt, F. T. M., and H. van Dop, 1982: Atmospheric Turbulence and Air Pollution Modeling. F.T.M. Nieuwstadt and H. van Dop, eds., D. Reidel Publishing Company, Dordrecht, Holland.

Ogawa, Y., P. G. Diosey, J. Uehara, and H. Ueda, 1985: Wind Tunnel Observations of Flow and Diffusion under Stable Stratification. Atmos. Environ., 19: 65-74.

Panofsky, H. A., and B. Prasad, 1965: Similarity Theories and Diffusion. Int. J. Air and Water Pollution, 9: 419-430.

Panofsky, H. H., H. Tennekes, D. H., Lenschow, and J. C. Wyngaard, 1977: The Characteristics of Turbulent Velocity Components in the Surface Layer under Convective Conditions. *Boundary-Layer Meteor.*,11: 355-361.

Pasquill, F., 1961: The Estimation of the Dispersion of Windborne Material, *Meteor. Mag.*,90: 33.

Pasquill, F., 1971: Atmospheric Dispersion of Pollution. *Quart. J. Roy. Meteor. Soc.*,97: 369-395.

Pasquill, F., 1974: *Atmospheric Diffusion*, 2nd Ed., Ellis Horwood Ltd., Chichester, England, 429 pp.

Pasquill, F., 1976a: *Atmospheric Dispersion Parameters in Gaussian Plume Modeling - Part II; Possible Requirements for Change in the Turner Workbook Values.* EPA-600/4-76-030b, U.S. Environmental Protection Agency, Research Triangle Park, NC, 44 pp.

Pasquill, 1976b: *The "Gaussian-Plume" Model with Limited Vertical Mixing.* EPA-600/4-76-042, U.S. Environmental Protection Agency, Research Triangle Park, NC, 13 pp.

Pasquill, F., 1978: *Atmospheric Dispersion Parameters in Plume Modeling.* EPA-600/4-76-042, U.S. Environmental Protection Agency, Research Triangle Park, NC, 13 pp.

Pasquill, F., and F. B. Smith, 1983: *Atmospheric Diffusion,* Third Edition. Ellis Horwood Ltd, Chichester, England, 437 pp.

Pooler, F., Jr., 1966: A Tracer Study of Dispersion over a City. *J. Air Pollut. Control Assoc.*, 16: 677-681.

Poreh, M., and J. E. Cermak, 1984: Wind Tunnel Simulation of Diffusion in a Convective Boundary Layer. *Boundary-Layer Meteor.*,30: 431-455.

Prahm, L. P., R. Berkowicz, and O. Christensen, 1979: Generalization of K-Theory for Turbulent Diffusion. Part II. Spectral Diffusivity Model for Plume Dispersion. *J. Appl. Meteor.*,18: 273-282.

Randerson, D., 1979: Review Panel on Sigma Computations. *Bull. Amer. Meteor. Soc.*,60: 682-683.

Randerson D., editor, 1984: *Atmospheric Science and Power Production.* U.S. Dept. of Energy, NTIS No. DE84005177, 850 pp.

Rao, K. S., and H. F. Snodgrass, 1978: *The Structure of the Nocturnal Boundary Layer.* NOAA, Air Resources Laboratory, Oak Ridge, TN, 44 pp.

Roberts, O. F. T., 1923: The Theoretical Scattering of Smoke in a Turbulent Atmosphere. *Proc. Roy. Soc.* A, 104: 640.

Robins, A. G., 1978: Plume Dispersion from Ground-Level Sources in Simulated Atmospheric Boundary Layers. *Atmos. Environ.*,12: 1033-1044.

Sagendorf, J. F., and C. R. Dickson, 1974: *Diffusion under Low Windspeed, Inversion Conditions.* NOAA TM ERL ARL-52, NOAA, Air Resources Laboratory, Idaho Falls, ID, 89 pp.

Sawford, B. L., 1982a: Comparison of Some Different Approximations in the Statistical Theory of Relative Dispersion. *Quart. J. Roy. Meteor. Soc.*,108: 191-206.

Sawford, B. L., 1982b: Lagrangian Monte Carlo Simulation of the Turbulent Motion of a Pair of Particles. *Quart. J. Roy. Meteor. Soc.*,108: 207-213.

Schiermeier, F. A., W. E. Wilson, F. Pooler, J. K. S. Ching, and J. F. Clarke, 1979: Sulfur Transport and Transformation in the Environment (STATE): A Major EPA Research Program. *Bull. Amer. Meteor. Soc.*, 60: 1303-1312.

Schiermeier, F. A., 1984: *Scientific Assessment Document of Status of Complex Terrain Dispersion Models for EPA Regulatory Applications.* EPA-600/3-84-103, U.S. Environmental Protection Agency, Atmospheric Sciences Research Laboratory, Research Triangle Park, NC, 132 pp.

Simpson, C. L., 1961: *Some Measurements of the Deposition of Matter and its Relation to Diffusion from a Continuous Point Source in a Stable Atmosphere.* Report NW-69292. Rev., Hanford Atomic Products Operation, Hanford, Wash.

Singer, I. A., and M. E. Smith, 1966: Atmospheric Dispersion at Brookhaven National Laboratory. *Int. J. Air and Water Pollut.* 10: 125-135.

Sivertsen, B., S. E. Gryning, A. A. Holtslag, and J. S. Irwin, 1985: Atmospheric Dispersion Modeling Based Upon Boundary Layer Parameterization. *15th Int. Tech. Meeting on Air Pollution and Its Application.* St. Louis, Missouri. 16 pp.

Sklarew, R. C., and A. Joncich, 1979: *Survey of Existing Data Bases For Plume Model Validation Studies.* Final Report, EPRI EA 1159 TPS 78-806, Electric Power Research Institute, Palo Alto, CA, pp .

Slade, D. H.,ed., 1968: *Meteorology and Atomic Energy*, U.S. Atomic Energy Commission, Germantown, MD, 446 pp.

Smagorinsky, J., 1963: General Circulation Experiments with the Primitive Equations:1 The Basic Experiment. *Mon. Wea. Rev.*,91: 99-164.

Smith, F. B., 1968: Conditioned Particle Motion in a Homogeneous Turbulent Field. *Atmos. Environ.*,2: 491-508.

Smith, F. B., 1972: A Scheme for Estimating the Vertical Dispersion of a Plume From a Source Near Ground Level. NATO/CCMS No. 14. In: Proceedings of the Third Meeting of the Expert Panel on Air Pollution Modeling, NATO/CCMS, Brussels, Belgium, pp. XVII-1 - XVII-14.

Smith, F. B., 1981: The Significance of Wet and Dry Synoptic Regions on Long-Range Transport of Pollution and its Deposition. Atmos. Environ., 15: 863.

Smith, F. B., 1983: Discussion of 'Horizontal Diffusion on the Atmosphere: A Lagrangian-Dynamical Theory' by F. A. Gifford, Atmos. Environ., 17: 194-197.

Smith, F. B., and J. S. Hay, 1961: The Expansion of Clusters of Particles in the Atmosphere. Quart. J. Roy. Meteor. Soc.,87: 82.

Smith, F. B., and D. Thomson, 1984: Solutions of the Integral Equation of Diffusion and the Random Walk Model for Continuous Plumes and Instantaneous Puffs in the Atmospheric Boundary Layer. Boundary-Layer Meteor.,30: 143-157.

Smith, M. E., 1956: The Variation of Effluent Concentrations from an Elevated Point Source. A. M. A. Arch. Indust. Health, 14: 56-68.

Smith, M. E., 1981: Transport and Diffusion Modeling and its Application. In: Air Quality Modeling and the Clean Air Act, Amer. Meteor. Soc., Boston, MA, pp. 228-270.

Snyder, W. H., 1972: Similarity Criteria for the Application of Fluid Models to the Study of Air Pollution Meteorology, Boundary-Layer Meteor.,3: 113-134.

Snyder, W. H., 1981: Guideline for Fluid Modeling of Atmospheric Diffusion. EPA 600/8-81-009, U.S. Environmental Protection Agency, Atmospheric Sciences Research Laboratory, Research Triangle Park, NC, 185 pp.

Strimaitis, D., G. Hoffnagle, and A. Bass, 1981: On-Site Meteorological Instrumentation Requirements to Characterize Diffusion from Point Sources: Workshop Report. EPA-600/9-81-020, U.S. Environmental Protection Agency, Research Triangle Park, NC, 124 pp.

Strimaitis, D., A. Venkatram, B. R. Greene, S. Hanna, S. Heisler, T. F. Lowery, A. Bass, and B. A. Egan, 1983: EPA Complex Terrain Model Development: Second Milestone Report - 1982, EPA-600/3-83-015, U.S. Environmental Protection Agency, Atmospheric Sciences Research Laboratory, Research Triangle Park, NC, 402 pp.

Stull, R. B., 1984: Transilient Turbulence Theory. Part I: The Concept of Eddy Mixing Across Finite Distances. J. Atmos. Sci., 41: 3351-3367.

Stull, R. B., 1985: A Fair Weather Cumulus Cloud Classification Scheme for Mixed-Layer Studies. J. Climate and Appl. Meteor., 24: 49-56.

Stull, R. B., and E. W. Eloranta, 1984: Boundary Layer Experiment - 1983. Bull. Amer. Meteor. Soc., 65: 450-456.

Stull, R. B., and T. Hasegawa, 1984: Transilient Turbulence Theory. Part II: Turbulent Adjustment. J. Atmos. Sci., 41: 3368-3379.

Sutton, O. G., 1953: Micrometeorology, McGraw-Hill, New York.

Taylor, G. I., 1921: Diffusion by Continuous Movements. Proc. London. Math. Soc., Ser. 2, 20: 196.

Tennekes, H., 1984: Discussion in Boundary-Layer Meteor., 30: 104.

Tennessee Valley Authority, 1970: Tennessee Valley Authority Environmental Research and Development Report on Full-Scale Study of Inversion Breakup at Large Power Plants, Muscle Shoals, AL.

Thomson, D. J., 1984: Random Walk Modeling of Diffusion in Inhomogeneous Turbulence. Quart. J. Roy. Meteor. Soc., 110: 1107-1120.

Thomas, P. W., and K. Nester, 1976: Experimental Determination of the Atmospheric Dispersion Parameters over Rough Terrain, Part II, Evaluation of Measurements. Rep. KFK2286, Central Division of Nuclear Research, M. B. H., Federal Republic of Germany, 116 pp.

Thomas, P. W., W. Hubschmann, L. A. Konig, H. Schuttelkopf, S. Vogt, and M. Winter, 1976: Experimental Determination of the Atmospheric Dispersion Parameters over Rough Terrain, Part I, Measurements at the Karlshruhe Nuclear Research Center, Report KFK2285, Central Division of Nuclear Research, M.B.H., Federal Republic of Germany, 132 pp.

Turner, D. B., 1961: Relationships Between 24-Hour Mean Air Quality Measurements and Meteorological Factors in Nashville, TN. J. Air Pollut. Control Assoc.,11: 483-489.

Turner, D. B., 1970: Workbook of Atmospheric Dispersion Estimates. U. S. Environmental Protection Agency, Atmospheric Sciences Research Laboratory, Research Triangle Park, NC, 84 pp.

Turner, D. B., and J. S. Irwin, 1982: Extreme Value Statistics Related to Performance of a Standard Air Quality Simulation Model Using Data at Seven Power Plant Sites. Atmos. Environ., 16: 1907-1914.

van Dop, H., F. T. M. Nieuwstadt, and J. C. R. Hunt, 1985: Random Walk Models for Particle Displacements in Inhomogeneous Unsteady Turbulent Flows. Phys. Fluids, 28: 1639-1653.

Van Duuran, H., and F. T. M. Nieuwstadt, 1980: Dispersion from the 213m High Meteorological Mast at Cabauw in the Netherlands. In: *Studies in Environmental Science*, 8. Elsevier, Amsterdam, The Netherlands, 77-90.

van Ulden, A. P., 1978: Simple Estimates for Vertical Diffusion from Sources Near the Ground. *Atmos. Environ*.,12: 2125-2129.

van Ulden, A. P., and A. A. M. Holtslag, 1983: The Stability of the Atmospheric Surface Layer During Nighttime. *Sixth Symposium on Turbulence and Diffusion*. Amer. Meteor. Soc. Boston, MA.,pp. 257-260.

van Ulden, A. P., and A. A. M. Holtslag, 1985: Estimation of Atmospheric Boundary Layer Parameters for Diffusion Applications. *J. Climate and Appl. Meteor*.,(in press).

Venkatram, A., 1980: Estimating the Monin-Obukhov Length in the Stable Boundary Layer for Dispersion Calculations. *Boundary-Layer Meteor*., 19: 481-485.

Venkatram, A., 1982: A Semi-Empirical Method to Compute Concentrations Associated with Surface Releases in the Stable Boundary Layer. *Atmos. Environ*.,16: 245-248.

Venkatram, A., D. Strimaitis, and D. Dicristofaro, 1984: A Semi-Empirical Model to Estimate Vertical Dispersion of Elevated Releases in the Stable Boundary Layer. *Atmos. Environ*.,18: 923-928.

Venkatram, A., D. Strimaitis, and W. Eberhard, 1983: Dispersion of Elevated Releases in the Stable Boundary Layer. *Sixth Symposium on Turbulence and Diffusion*. Amer. Meteor. Soc. Boston, MA, pp. 297-299.

Vogt, K. J., and H. Geis, 1974: *Tracer Experiments on the Dispersion of Plumes over Terrain of Major Surface Roughness*, Report Jul-113, Central Division for Radiation Protection, Kernforshungsanlage Julich GmbH, Julich, Federal Republic of Germany.

Voloshchuk, V. M., 1976: Nonlocal Parameterization of Vertical Turbulent Exchange in the Surface Layer. *Meteorologiva i Gidrologivia*, 6: 35-42.

Walker, E. R., 1965: A Particulate Diffusion Experiment. *J. Appl. Meteor*., 4, 614-621.

Weber, A. H., 1976: *Atmospheric Dispersion Parameters in Gaussian Plume Modeling: Part 1 - Review of Current Systems and Possible Future Developments*. EPA-600/4-76-030a. U.S. Environmental Protection Agency, Research Triangle Park, NC, 59 pp.

Weber, A. H., K. R. McDonald, and G. A. Briggs, 1977: Turbulence Classification Schemes for Stable and Unstable Conditions. *Joint Conference on Applications on Air Pollution Meteorology*. Amer. Meteor. Soc. Boston, MA, pp. 96-102.

Weil, J. C., 1977: *Evaluation of the Gaussian Plume Model at Maryland Power Plants*. PPSP-MP-16. Maryland Power Plant Siting Program, Department of Natural Resources, Annapolis, MD.

Weil, J. C., 1979: *Assessment of Plume Rise and Dispersion Models Using Lidar Data*. PPSP-MP-24. Maryland Power Plant Siting Program. Department of Natural Resources, Annapolis, MD.

Weil, J. C., 1983: Application of Advances in Planetary Boundary Layer Understanding to Diffusion Modeling. *Sixth Symposium on Turbulence and Diffusion*. Amer. Meteor. Soc., Boston, MA, pp. 42-46.

Weil, J. C., 1985: Updating Applied Diffusion Models. *J. Climate. Appl. Meteor*. (in press).

Weil, J. C., and R. P. Brower, 1982: *The Maryland PPSP Dispersion Model for Tall Stacks.* PPSP-MP-36. Maryland Power Plant Siting Program. Department of Natural Resources, Annapolis, MD.

Weil, J. C., and R. P. Brower, 1984: An Updated Gaussian Plume Model for Tall Stacks. *J. Air Pollut. Control Assoc.* 34: 818-827.

Weiringa, J., 1976: An Objective Exposure Correction Method for Average Wind Speeds Measured at a Sheltered Location. *Quart. J. Roy. Meteor. Soc*., 102: 241-253.

Wieringa, J., 1980: Representativeness of Wind Observations at Airports. *Bull. Amer. Meteor. Soc*., 51: 962-971

Wetzel, P. J., 1982: Toward Parameterization of the Stable Boundary Layer. *J. Appl. Meteor*.,21: 7-13.

Wilde, N. P., R. B. stull, and E. W. Eloranta, 1985: The LCL and Cumulus Onset. *J. Climate and Appl. Meteor*.,24: 640-657.

Willis, G. E., and J. W. Deardorff, 1974: A Laboratory Model of the Unstable Planetary Boundary Layer. *J. Atmos. Sci*.,31: 1297-1307.

Willis, G. E., and J. W. Deardorff, 1976a: A Laboratory Model of Diffusion into the Convective Planetary Boundary Layer. *Quart. J. Roy. Meteor. Soc*.,102: 427-445.

Willis, G. E., and J. W. Deardorff, 19876b: Visual Observations of Horizontal Planiforms of Penetrative Convection. *Third Symposium on Atmospheric Turbulence, Diffusion, and Air Quality*. Amer. Meteor. Soc., Boston, MA, pp. 9-12.

Willis, G. E., and J. W. Deardorff, 1978: A Laboratory Study of Dispersion from an Elevated Source in a Convective Mixed Layer. *Atmos. Environ*., 12: 1305-1313.

Willis, G. E., and J. W. Deardorff, 1981: A Laboratory Study of Dispersion from a Source in the Middle of the Convectively Mixed Layer. Atmos. Environ.,15: 109-117.

Willis, G. E., and J. W. Deardorff, 1983: On Plume Rise Within a Convective Boundary Layer. Atmos. Environ., 17: 2435-2447.

Willis, G. E., and N. Hukari, 1984: Laboratory Modeling of Buoyant Stack Emissions in the Convective Boundary Layer. Fourth Joint Conference on Applications of Air Pollution Meteorology. Amer. Meteor. Soc., Boston, MA, pp. 24-26.

Wilson, D. J., A. G. Robbins, and J. E. Fackrell, 1982a: Predicting the Spatial Distribution of Concentration Fluctuations from a Ground-Level Source. Atmos. Environ.,16: 497-504.

Wyngaard, J. C., 1975: Modeling the Planetary Boundary Layer - Extension to the Stable Case. Boundary-Layer Meteor.,9: 441-460.

Wyngaard, J. C., O. R. Kote, and K. S. Rao, 1974: Modeling the Atmospheric Boundary Layer. In: Advances in Geophysics, Vol. 18A: 193-211.

Wyngaard, J. C., and R. A. Brost, 1984: Top-Down and Bottom-Up Diffusion of a Scalar in the Convective Boundary Layer. J. Atmos. Sci., 41: 102-112.

Yamada, T., and G. Mellor, 1975: A Simulation of the Wangara Atmospheric Boundary Layer. J. Atmos. Sci., 32: 2309-2329.

Yokoyama, O., M. Gamo, and S. Yamamoto, 1977: On the Turbulence Quantities in the Neutral Atmospheric Boundary Layer. J. Meteor. Soc. Japan, 55: 312-318.

Zhou, L., and H. A. Panofsky, 1983: Wind Fluctuations in Stable Air at the Boulder Tower. Boundary-Layer Meteor., 25: 353-362.

Zilitinkevich, S. S., 1972: On the Determination of the Height of the Ekman Boundary Layer. Boundary-Layer Meteor., 3: 141-145.